APPENDICES

Appendix A
Results of aquifer testing within the study area

AQUIFER TEST ANALYSES

As part of this study, aquifer tests were done to evaluate the hydraulic properties of the Navajo Sandstone aquifer and the Pine Valley monzonite aquifer. Four aquifer tests were done in the Navajo Sandstone and one aquifer test was done in the Pine Valley monzonite. The locations of these five tests are shown in figure A-1.

Hurricane Bench Aquifer Test

The purpose of the Hurricane Bench aquifer test was to determine the transmissivity and storage properties of the Navajo aquifer about 5 mi southwest of Hurricane, Washington County, Utah (fig. A-2). The aquifer test was conducted in January and February 1996 by the USGS in coordination with the Winding Rivers Corpo-

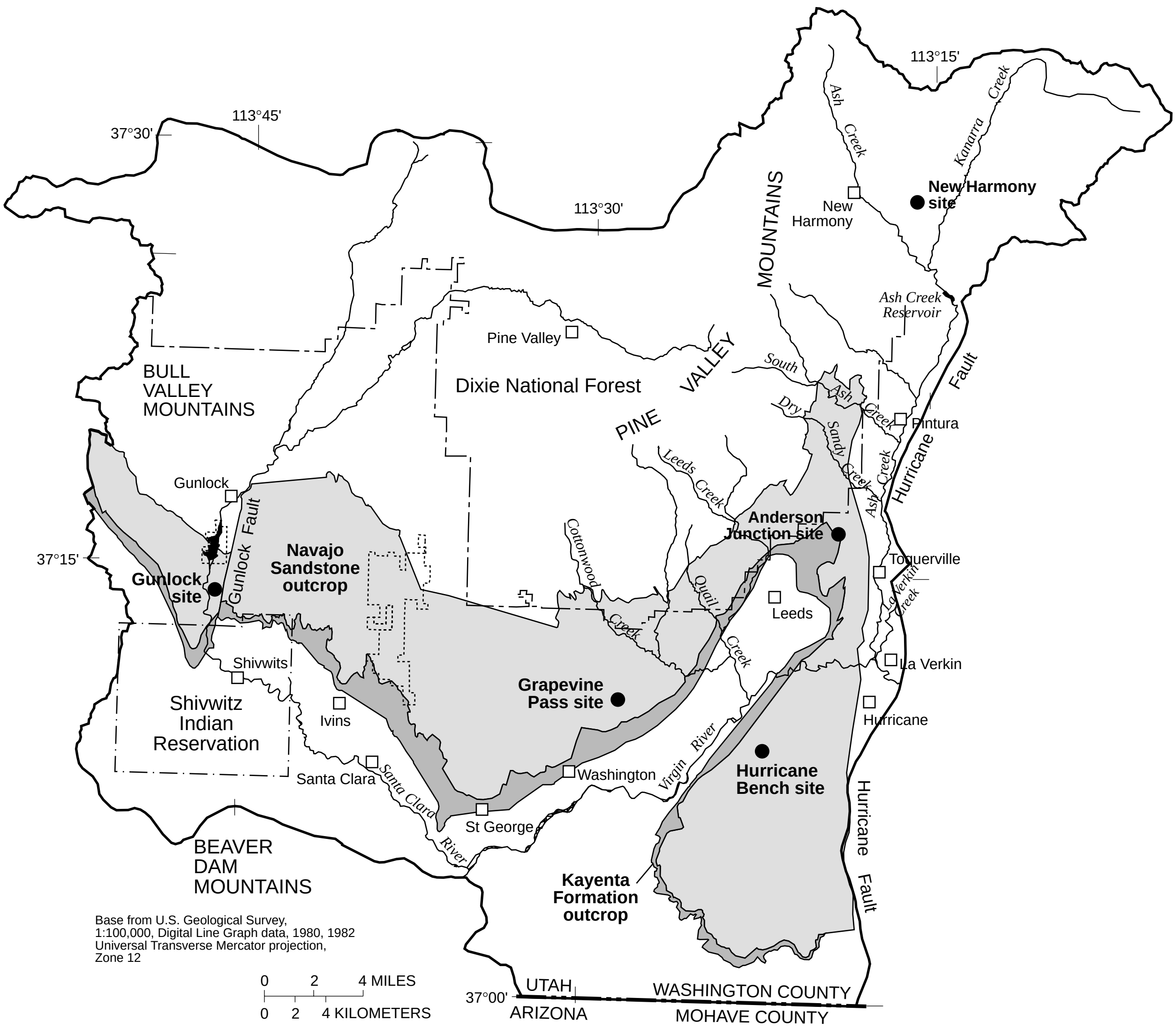


Figure A-1. Location of aquifer-test sites within the central Virgin River basin study area, Washington County, Utah, 1996.

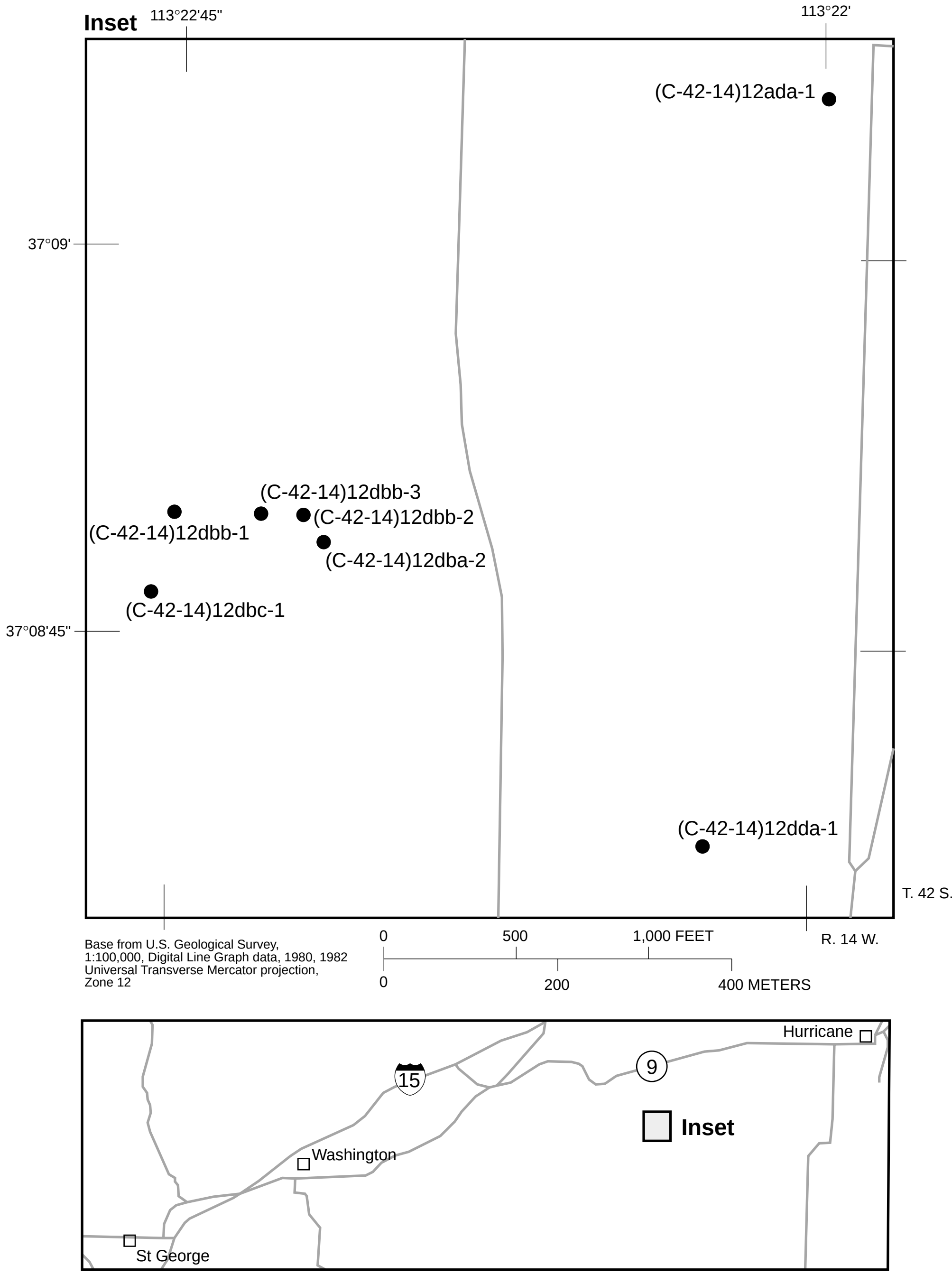


Figure A-2. Location of wells in the Hurricane Bench aquifer test, Washington County, Utah, January and February 1996.

ration. The multiple-well aquifer test involved pumping well (C-42-14)12dbb-2 at a rate of about 330 gal/min for almost 5 days. The discharge from the pumped well was diverted into an irrigation pipe, which transported the water more than 1 mi away from the well, where it was applied as irrigation water via sprinklers. Discharge was measured with a Clampatron flow meter, which was field checked by capturing discharge from one of the irrigation sprinkler nozzles and multiplying this by the total number of active nozzles. Because of problems with the circuit breaker controlling the submersible pump, some pumping occurred during the 2 days prior to the aquifer test. Therefore, only data from the recovery part of this aquifer test were analyzed.

Analysis of geologic maps and drillers' logs indicates that the Navajo Sandstone aquifer is areally extensive at the aquifer-test site and is underlain by the less permeable Kayenta Formation. Information from the drillers' log of well (C-42-13)7bcc-3 (0.45 mi from the pumping well) indicates that sandstone was found from 12 to 1,450 ft below land surface; and alternating siltstone and sandstone layers characteristic of the Kayenta Formation were found from 1,450 ft to 1,860 ft. Because of the shallow dip of the Navajo Sandstone to the northeast and the average prepumping depth to water of about 60 ft, the average saturated thickness of the Navajo aquifer at the aquifer-test site was assumed to be about 1,350 ft. According to a recent Utah Geological Survey fracture study of a nearby Navajo outcrop on Sand Mountain (about 3 1/2 mi to the south), the predominant fracture direction is northeast-southwest and the secondary direction is east-west.

Water levels were measured in six observation wells and the pumped well for 6 days preceding the test, during the 7 days of pumping, and for 6 days after the pump was shut off. Data for the pumped well and observation wells are recorded in table A-1. Drawdown and recovery of sufficient magnitude to analyze were observed in five of the six observation wells. There was no noticeable drawdown or recovery at the observation well farthest from the pumped well, (C-42-14)12ada-1. All of the observation wells are finished in the Navajo Sandstone aquifer. The two observation wells nearest the pumped well have similar perforation intervals to the pumped well. The four observation wells farther away generally are open to the aquifer at shallower depths than the pumped well.

The measured water levels at five observation wells were corrected for barometric changes assuming 100 percent barometric efficiency. This barometric efficiency was chosen on the basis of observations of prepumping water-level changes at observation well (C-42-14)12dba-2 as a result of changes in barometric pressure. The 100-percent correction was verified by a comparison of the effects on water levels of barometric efficiencies ranging from 50 to 100 percent. Barometric data from a mercury barometer located at the Cedar City Airport, about 30 mi to the north, was used for this correction.

As mentioned above, only the recovery data from the aquifer test was analyzed. Because water levels in the affected observation wells did not reach a pumping equilibrium before the pump was shut off, the recovery data were affected by residual drawdown and trend corrections to the recovery data were necessary. Straight-line fits to semilogarithmic plots of the water levels in the observation wells during drawdown were used for determining the prerecovery trend. This prerecovery

Table A-1. Construction data for the wells used in the Hurricane Bench aquifer test, Washington County, Utah

[N/A, not applicable]

Well number	Radial distance (feet)	Casing diameter (inches to feet)	Open interval (feet below land surface)	Opening type
(C-42-14)12dbb-2	N/A	12 to 560	62 - 560	Screen
(C-42-14)12dba-2	34	12 to 510	120 - 510	Perforations
(C-42-14)12dbb-3	106	12 to 510	58 - 510	Screen
(C-42-14)12dbb-1	475	12 to 23	23 - 140	Open hole
(C-42-14)12dbc-1	590	10 to 100	100 - 270	Open hole
(C-42-14)12dda-1	1,665	12 to 40	40 - 425	Open hole
(C-42-14)12ada-1	2,500	12 to 300	101 - 300	Open hole

trend was extended through the recovery part of the aquifer test to correct for this trend.

Recovery data for the observation wells were initially plotted together by dividing time in elapsed minutes by the observation well's radial distance squared. However, recovery data from the closest observation well, (C-42-14)12dba-2, were eliminated from the analysis because its initially steep response is assumed to be affected by well-bore storage because its proximity to the pumping well (34 ft) and its large borehole size (12-in. diameter). Also, the maximum drawdown and recovery at this observation well were a substantial part of the saturated thickness of the aquifer and would cause the transmissivity to change during the aquifer test. Recovery data from observation well (C-42-14)12dda-1 were also eliminated because the barometric pressure and prerecovery-trend corrections were a substantial part of the recovery (as much as 0.5 ft of the total 1.6 ft of recovery) and could have introduced error into the analysis. Therefore, only recovery data from observation wells (C-42-14)12dbb-3, (C-42-14)12dbb-1, and (C-42-14)12dbc-1, were analyzed. Because the maximum 50-ft drawdown and recovery measured at the closest of the three wells, (C-42-14)12dbb-3), was less than 4 percent of the saturated thickness, changes in transmissivity with change in the saturated thickness of the aquifer were not substantial at these three observation wells. Because these three observation wells are all at a similar radial orientation with respect to the pumping well, the degree of horizontal anisotropy resulting from fractures within the sandstone could not be evaluated.

The Theis solution (1935) for confined aquifers was first chosen for the analysis. The aquifer is assumed to act as confined, as indicated by (1) the high barometric efficiency observed at well (C-42-14)12dba-2, and (2) drillers' logs for wells (C-42-14)12dbb-1 and (C-42-14)12dda-1, drilled with a cable-tool rig, both indicate that water was initially encountered deeper (12 ft and 6 ft, respectively) than the static water levels later measured in the wells. The Theis method assumes that water is released instantaneously from storage with a decline of hydraulic head. However, scatter in the composite plot of recovery versus time for these three observation wells indicated that well responses varied substantially from the Theis-type response. Calculations of transmissivity and storage values for the individual observation wells were also determined from separate time-recovery plots for each well, either using the Cooper-Jacob solution (Cooper and Jacob, 1946) or the Theis solution. These calculations show generally higher values of transmissivity with increased distance from the pumped well. Assuming that the Navajo aquifer in this region is fairly homogeneous, this apparent increase in transmissivity with radial distance may indicate the effects of leakage or delayed yield. Also, the confined-type response to barometric pressure changes may only indicate the aquifer's confined response to small stresses; larger stresses may result in dewatering and a conversion to an unconfined aquifer response at closer observation wells.

Therefore, the recovery data were reanalyzed with the modified Hantush solution (Lohman, 1972, p. 32-34) for leaky confined aquifers with vertical movement. This solution provided the best fit to the composite plot of recovery data from the three observation wells (fig. A-3). This method was chosen because of the possibility that the underlying Kayenta Formation may act as a low-permeability layer and release a relatively large amount of water from storage as a result of pumping in the overlying Navajo aquifer. Hydraulic conductivity can be calculated by dividing transmissivity either by saturated thickness of the aquifer by the saturated thickness of the perforated interval of the production well. Transmissivity and storage-coefficient values calculated with this method are 1,075 ft^2/d and 0.002, respectively. Assuming an average saturated aquifer thickness of 1,350 ft, the calculated hydraulic conductivity is 0.8 ft/day. This hydraulic conductivity value of 0.8 ft/d is smaller than the average value of 2.1 ft/day determined from laboratory analyses of outcrop samples of the Navajo Sandstone (Cordova, 1978). However, dividing the transmissivity by the 500-ft perforated interval of the production well results in a hydraulic conductivity of 2.2 ft/day, which is similar to the average value. This larger value is preferred because it is likely that small bedding plane features, such as thin, finer-grained layers formed during sand-dune deposition, reduce vertical hydraulic conductivity and vertical flow to the well from the deeper, unpenetrated part of the aquifer.

Anderson Junction Aquifer Test

The purpose of the Anderson Junction aquifer test was to determine the transmissivity and storage properties of the Navajo aquifer near Anderson Junction in Washington County, Utah (fig. A-4). The aquifer test was conducted in March and April 1996 by the USGS in coordination with the Washington County Water Conservancy District. The multiple-well aquifer

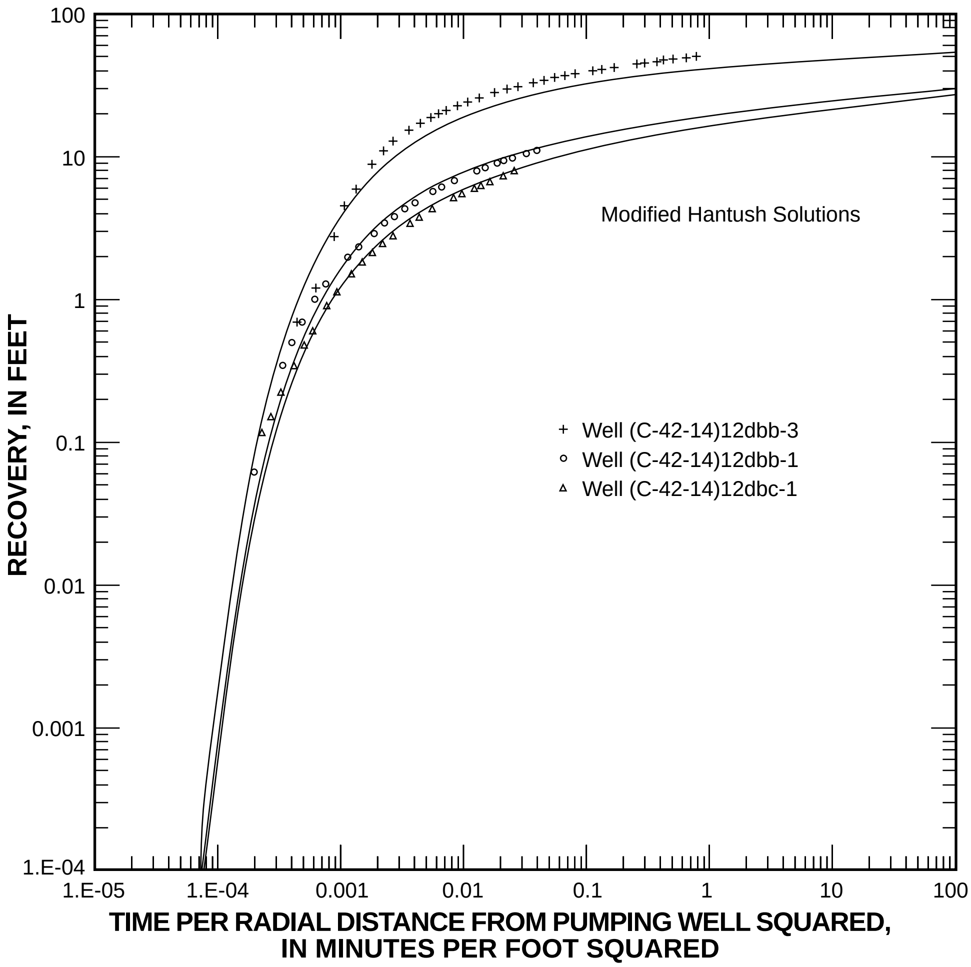
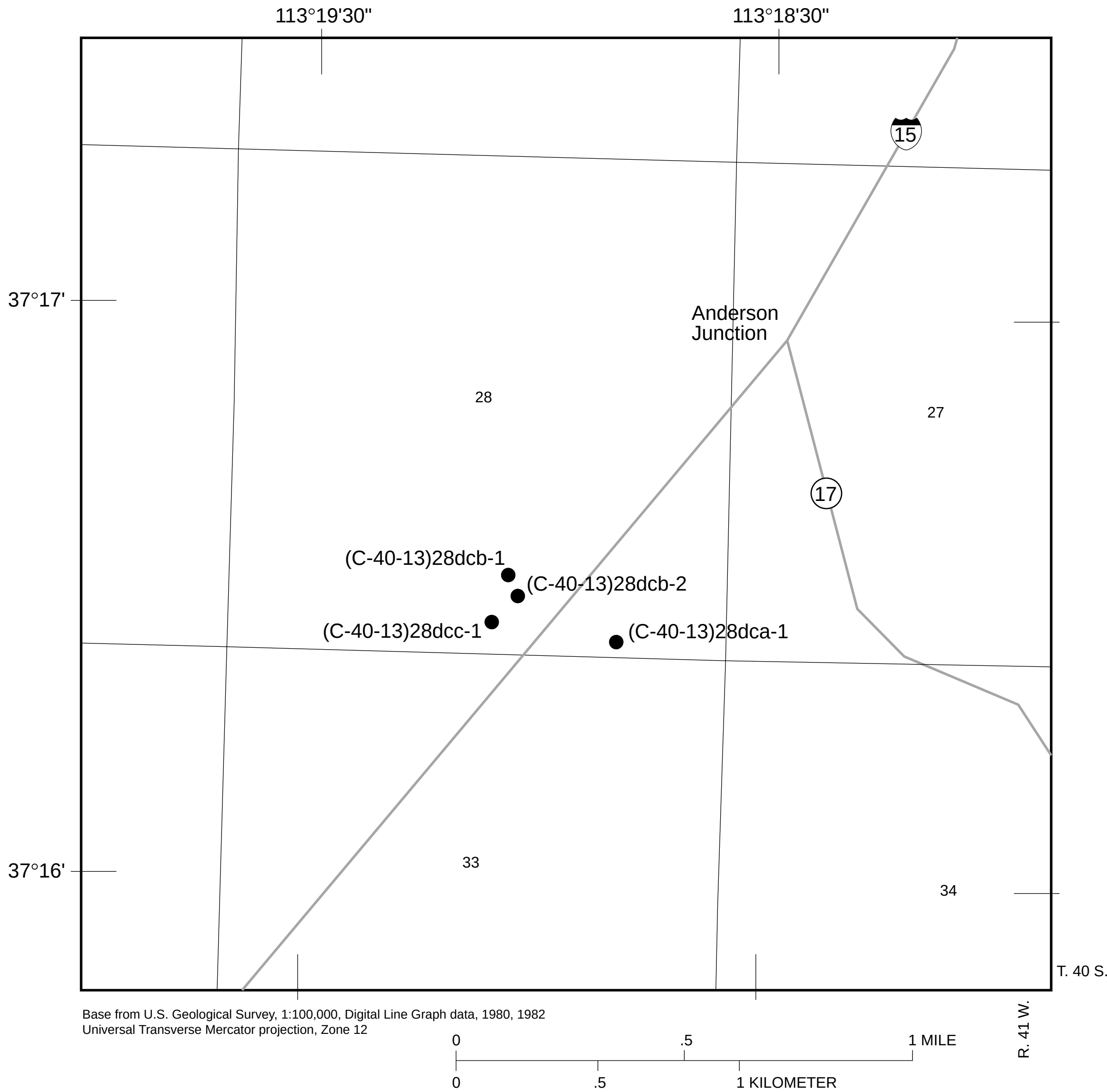


Figure A-3. Recovery data from wells during the Hurricane Bench aquifer test, Washington County, Utah, January and February 1996 (modified Hantush solution (Lohman, 1972)).

test involved pumping at well (C-40-13)28dcb-2 for about 4 days at an average rate of 1,100 gal/min. Discharge was measured with a pito tube, v-notch weir, and pygmy meter. The discharge from the production well was diverted into a 15-in. diameter ABS drain pipe, which transported the water 500 ft away from the well to a natural dry wash heading southeast under Highway I-15. Water levels were measured in three observation wells and the production well, (C-40-13)28dcb-2, for 4 days prior to the test, during the 4 days of pumping, and for as many as 20 days after the pump was shut off.

The aquifer-test site is in a highly fractured region of the Navajo Sandstone outcrop that has two predominant clusters of fracturing at orientations of 180 to 210 degrees and 90 to 130 degrees (Hurlow, 1998). On the basis of the Utah Geological Survey's fracture study, two observation wells were drilled specifically for the aquifer test at approximately the same radial distance from the pumped well, but at perpendicular orientations. Well (C-40-13)28dca-1, herein referred to as well A, is located 383 ft east-southeast of the production well along a 110-degree orientation (parallel to the 90 to 130 azimuthal cluster of fractures). Well (C-40-13)28dcc-1, herein referred to as well B, is located 376 ft south-southwest of the production well along a 200-degree orientation (parallel to the 180 to 210 degree azimuthal cluster of fractures). Well (C-40-13)28dcb-1, herein referred to as the original well, is located 10 ft due east of the pumped well. Data for the pumped well and observation wells are recorded in table A-2.

According to conversations with the driller and information from the drillers' logs, the Navajo Sandstone aquifer. Because wells A and B were both drilled by using the reverse rotary method with air, the driller



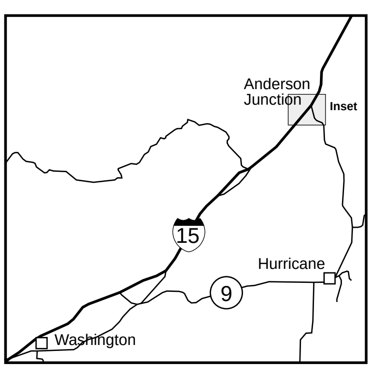


Figure A-4. Location of wells in the Anderson Junction aquifer test, Washington County, Utah, March and April 1996.

Table A-2. Construction data for the wells used in the Anderson Junction aquifer test, Washington County, Utah

Well name	Well number	Radial distance (feet)	Casing diameter (inches to feet)	Open interval (feet below land surface)	Opening type
Pumped well	(C-40-13)28dcb-2	0	16 to 500	110 - 470	Screen
Original well	(C-40-13)28dcb-1	10	6 to 225	160 - 225	Perforations
Well B	(C-40-13)28dcc-1	376	5 to 400	160 - 380	Perforations
Well A	(C-40-13)28dca-1	383	5 to 400	160 - 380	Perforations

could readily identify when the water table was reached (the pumped well was drilled with water, so such a determination could not be made). According to the driller, water was found in well B at a depth of 190 ft and afterwards rose in the casing to a depth of 31 ft. In well A, water was found at a depth of 56 ft and rose in the casing to a depth of 21 ft. The potentiometric surface is nearly flat between the two wells. According to the drillers' logs, the lithology causing the confined conditions is different at two of the observation wells. At well B, vertical anisotropy within the Navajo Sandstone (possibly as a result of grain alignment, cementation, or finer sediments) appears to cause the confined conditions. At well A, the confined conditions are probably caused by a poorly permeable layer of clays, silts, and sands in the unconsolidated alluvium overlying the Navajo Sandstone.

Measured water levels at the observation wells were not corrected for barometric changes because the magnitude of drawdown and recovery at all of the wells was much larger (19 to 80 ft) than the effects of barometric changes (generally less than 1 ft). Prepumping trend corrections were applied to all of the observation-well drawdown data because of a rise in water levels resulting from recovery after the development of the production well shortly before the aquifer test. Prerecovery trend corrections were applied to the observation-well recovery data because water levels did not reach a pumping equilibrium before the pumping well was shut off on March 22, 1996.

The drawdown and recovery data for the three observation wells were initially plotted together on log-log scale by dividing time by the observation well's radial distance squared. The drawdown and recovery data from the closest observation well (original well) were eliminated from the analysis because of delayed response in early time data caused by well-bore storage effects resulting from proximity to the pumped well and large borehole size. Also, the maximum drawdown and recovery at this observation well (as much as 80 ft) made up a substantial part of the saturated thickness of the aquifer and would result in a substantial change in transmissivity during the aquifer test.

The data sets from the remaining two observation wells (wells A and B) were analyzed with three curve-matching solutions: (1) the Theis solution (1935) for confined aquifers, (2) the modified Hantush solution (Lohman, 1972, p. 32-34) for leaky confined aquifers with vertical movement, and (3) the Neuman solution (1974) for unconfined aquifers with delayed yield. None of these type curves fit both sets of data, indicating that the previously mentioned methods might not be applicable. In particular, the assumption of isotropy in the three methods is questionable. The presence of anisotropy at the Anderson Junction test site is indicated by the large difference in observed drawdown at the two observation wells: 33 feet at well A aligned with the 110-degree fracture orientation, and 19 ft at well B aligned with the 200-degree fracture orientation. These observations are consistent with a fractured anisotropic aquifer.

Therefore, a modified (simplified) version of a method presented by Papadopulos (1965) for data analysis from a homogeneous and anisotropic aquifer was used. The Papadopulos method assumes that the orientations of the principal axes directions for the transmissivity tensor are unknown. For the Anderson Junction aquifer test, the assumption is made that the two observation wells in the 110-degree and 200-degree orientations are parallel to the two principal axes.

Theory

The modification to the Papadopulos method was developed by Dr. Paul Hsieh of the USGS (written commun., 1997). Assuming that observation wells A and B are located along the maximum and minimum principal axes directions in an orthogonal orientation with respect to each other and the pumping well (fig. A-5), the drawdown in the observation wells are given by Papadopulos (1965, eq. 15 and 16):

$$s(x, y, t) = \frac{Q}{4\pi\sqrt{T_{xx}T_{yy}}} \cdot W(u_{xy}) \tag{A6}$$

with

$$u_{xy} = \frac{S}{4t} \cdot \left(\frac{T_{xx}y^2 + T_{yy}x^2}{T_{xx}T_{yy}}\right) \tag{A7}$$

where s is drawdown,
Q is pumping rate,
T_{xx} and T_{yy} are the transmissivities along principal axes,
S is aquifer storage,
t is time, and
$W(u_{xy})$ is the well function of u_{xy}.
Note that x and y here stand for ξ and η as described by Papadopulos (1965).

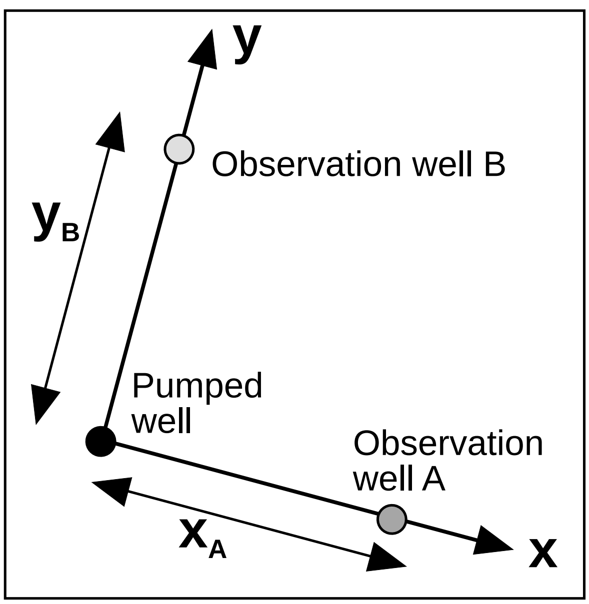


Figure A-5. Well-location geometry needed for applying modified version of Papadopulos solution (1965).

Applying the preceding solution to observation well A, which is located at $x = x_A$, $y = 0$ yields:

$$s(x_A, 0, t) = \frac{Q}{4\pi\sqrt{T_{xx}T_{yy}}} \cdot W\left(\frac{Sx_A^2}{4T_{xx}t}\right) \tag{A8}$$

In comparing this equation to the Theis solution:

$$s(r, t) = \frac{Q}{4\pi T} \cdot W\left(\frac{Sr^2}{4Tt}\right) \tag{A9}$$

the analogies are: T of the Theis equation is substituted with $\sqrt{T_{xx}T_{yy}}$, S/T of the Theis solution is substituted with S/T_{xx}, and r of the Theis equation is substituted with x_A. This analogy can be extended to the Cooper-Jacob straight-line method (Cooper and Jacob, 1946), which also can be modified for anisotropic conditions. Under ideal conditions in an anisotropic aquifer, Papadopulos shows that the straight-line parts of all observation well data on a semilog plot should have the same slope under ideal homogeneous conditions, so that the intercepts would yield T_{xx} greater than T_{yy} (fig. A-6). In the Cooper-Jacob method, the slope of the late time (straight-line part) data yields transmissivity from the determination of the change in drawdown per log cycle (Δs), and the intercept gives S/T, and thus S. Substituting $\sqrt{T_{xx}T_{yy}}$ for T yields the following equation modified from the Cooper-Jacob method, equations 5 and 6:

$$\sqrt{T_{xx}T_{yy}} = \frac{264(Q)}{(\Delta s)(7.48)} \tag{A10}$$

for T in ft^2/d, Q in gallons per minute, and Δs in feet. Likewise, substituting T_{xx}/S for T/S yields the following equation modified from the Cooper-Jacob method:

$$\frac{S}{T_{xx}} = \frac{2.25t_{0a}}{[r_a]^2} \tag{A11}$$

where t_{0a} is the x-intercept (time) for well A, and r_a is the radial distance to well A.
Next, applying the anisotropic solution to observation well B, which is located at $x = 0$, $y = y_B$, yields:

$$s(0, y_B, t) = \frac{Q}{4\pi\sqrt{T_{xx}T_{yy}}} \cdot W\left[\frac{Sy_B^2}{4T_{yy}t}\right] \tag{A12}$$

where $\sqrt{T_{xx}T_{yy}}$ is like T of Theis, S/T_{yy} is like S/T of Theis, and y_B is like r of Theis. After plotting the data from observation well B on semilog paper, the straight-line parts fitted to the data must have the same slope (and Δs) as the observation well A data set. This ensures that $T_{xx}T_{yy}$ computed from observation well B

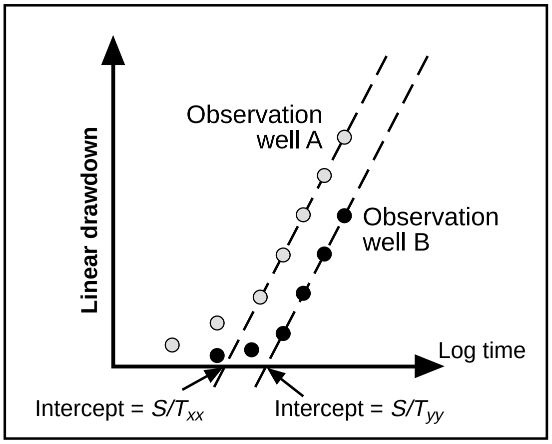


Figure A-6. Idealized data set for an anisotropic and homogeneous aquifer.

data is equal to $T_{xx}T_{yy}$ computed from observation well A data. By the same reasoning as above, substituting T_{yy}/S for T/S yields the following equation modified from the Cooper-Jacob method (Cooper and Jacob, 1946, eq. 8):

$$\frac{S}{T_{yy}} = \frac{2.25t_{0b}}{[r_b]^2} \qquad \text{(A13)}$$

where t_{0b} is the x-intercept (time) for well B, and r_b is the radial distance to well B.

In summary, the above straight-line fits to the observation well A and B data sets on a semilog graph should yield $\sqrt{T_{xx}T_{yy}}$ (the value from each of the data sets should be the same), S/T_{xx}, and S/T_{yy}. The following procedure can be used to determine Txx, Tyy, and S separately:

1. Square the $\sqrt{T_{xx}T_{yy}}$ to obtain $T_{xx}T_{yy}$.
2. Multiply S/T_{xx} and S/T_{yy} to obtain $S^2/(T_{xx}T_{yy})$
3. Multiply the result from steps 1 and 2 above to get S.
4. Divide the S obtained from Step 3 by S/T_{xx} to get Txx.
5. Divide the S obtained from Step 3 by S/T_{yy} to get T_{yy}.

T_{xx} is known as the "principal transmissivity in the direction of the x axis." T_{yy} is known as the "principal transmissivity in the direction of the y axis." If T_{xx} is greater than T_{yy}, then the x axis points along the major principal direction, and the y axis points along the minor principal direction. If T_{yy} is greater than T_{xx}, then the y axis points along the major principal direction, and the x axis points along the minor principal direction. Therefore, it is not necessary (nor warranted) to assume which is the major and which is the minor principal direction at the start of the analysis.

Application

With this modified version of the Papadopulos method, the corrected recovery data for both the south and east observation wells are plotted on a semilog graph. In an ideal homogeneous anisotropic aquifer, the slopes of observation-well data sets should be the same. However, unlike the ideal case, the slopes of the straight-line parts of the two observation-well data sets for this aquifer test are not identical (fig. A-7). With these two unequal slopes, the square root of $T_{xx}T_{yy}$ computed from well A does not equal that computed from well B. This indicates that the aquifer is not completely homogeneous at this site. Nonetheless, because the late-time data of each plot are similar and approach straight lines, the same slope was fitted to each data set. By forcing both lines to have the same slope the product of $T_{xx}T_{yy}$ from both wells is the same and the data can be interpreted using a homogeneous anisotropic aquifer model.

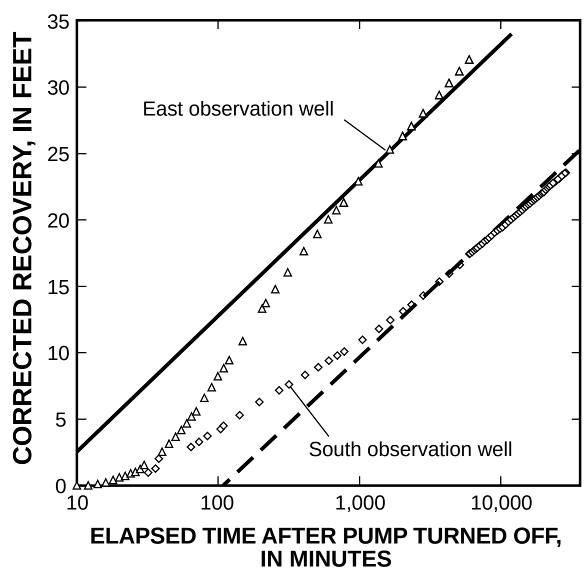


Figure A-7. Recovery data from wells during the Anderson Junction aquifer test, Washington County, Utah, March and April 1996.

The two fitted lines in figure A-7 have equal slopes (Δs) of 10 ft of drawdown per log cycle of time. Substituting these values into the Cooper-Jacob equation (5) where Q = 1,100 gal/min yields the relation:

$$\sqrt{T_{xx}T_{yy}} = 3,880\left(\frac{ft^2}{d}\right) \qquad \text{(A14)}$$

Also from figure A-7, the x-intercept on the semi-log plot for the well A recovery data is 5.5 minutes (0.0038 days); the x-intercept on the semi-log plot for the well B recovery data is 110.0 minutes (0.0764 days). Substituting the radial distance (r_a) to well A is 383 ft, and the radial distance (r_b) to well B is 376 ft into equations (6) and (8) yields:

$$\frac{S}{T_{xx}} = 5.829x10^{-8} \qquad \frac{S}{T_{yy}} = 1.216x10^{-6} \quad \text{(A15)}$$

Solving these three relation simultaneously yields $T_{xx} \cong 18,000$ ft^2/d, $T_{yy} \cong 900$ ft^2/d, and $S \cong 0.001$.

However, because heterogeneities within the Navajo aquifer at Anderson Junction do not permit a unique equal-slope fit to the semilog plot of observation-well data from wells A and B, an analysis of the possible range of values is necessary. To determine the maximum amount of interpretative error that may introduced by "forcing" lines of equal slope to both observation-well data sets, the steepest and shallowest possible fitted slopes are shown in figure A-8. The steepest possible slope for the two data sets corresponds to the best fit for the well A data set. The shallowest possible slope for the two data sets corresponds to the best fit for the well B data set. On the basis of these alternative slopes and x-intercepts, the range of values for T_{xx} ranges from 15,000 to 22,500 ft^2/d, T_{yy} from 650 to 900 ft^2/d, and S from 0.0007 to 0.0025. Therefore, the average of the maximum and minimum possible values for the transmissivity and storage from the Anderson Junction aquifer test, including error brackets, is $T_{xx} \cong 19,000$ ft^2/d ± 21%, $T_{yy} \cong 800$ ft^2/d ± 19%, and $S \cong 0.0013$ ± 1/4 log cycle. This indicates that the ratio of transmissivity (anisotropy factor) in the 110-degree and 200-degree orientations is about 24:1, but could range from 23:1 to 25:1, depending on the fitted slope. With an assumed aquifer thickness of 600 ft, horizontal hydraulic conductivity ranges from about 32 ft/d in the 110-degree orientation to 1.3 ft/d in the 200-degree orientation.

The range of hydraulic-conductivity values determined from this aquifer-test analysis is generally larger than Cordova's (1978, p. 26) laboratory determination of horizontal hydraulic-conductivity values that ranged from 0.36 to 5.0 ft/d. However, the laboratory-determined values do not include the effects of open fractures or other secondary openings that would increase the actual in-situ hydraulic conductivity.

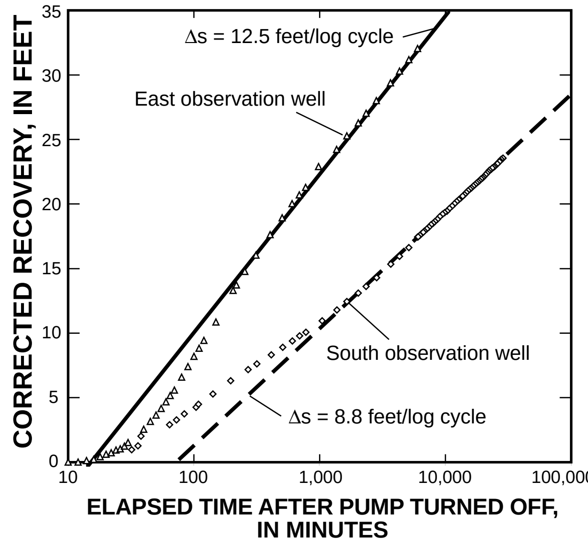


Figure A-8. Recovery data for wells during the Anderson Junction aquifer test showing range of possible slope, Washington County, Utah, March and April 1996.

Therefore, the Anderson Junction aquifer-test data may indicate that along the minor principal direction (200-degree orientation), the hydraulic-conductivity value of 1.3 ft/d is characteristic of unfractured rock and that the fractures along this orientation might be closed or unconnected. In the major principal direction (110-degree orientation), the hydraulic-conductivity value of 32 ft/d is about one order of magnitude higher than the range of laboratory values, indicating that fractures along this orientation might be open and more hydraulically connected.

Gunlock Well Field Aquifer Test

The purpose of the Gunlock Well Field aquifer test was to determine the transmissivity and storage properties of the Navajo aquifer downstream from the Gunlock Reservoir in Washington County, Utah (fig. A-9). The aquifer test was conducted in February 1996 by the USGS in coordination with the St. George Water and Power Department. The multiple-well aquifer test involved pumping at Gunlock well 7 for about 6 days at an average rate of 845 gal/min. Discharge was measured with an in-line flow meter. The discharge from the production well was diverted into a culinary supply line and removed from the aquifer-test site.

Water levels were measured in seven observation wells and the pumped well for about 18 days prior to

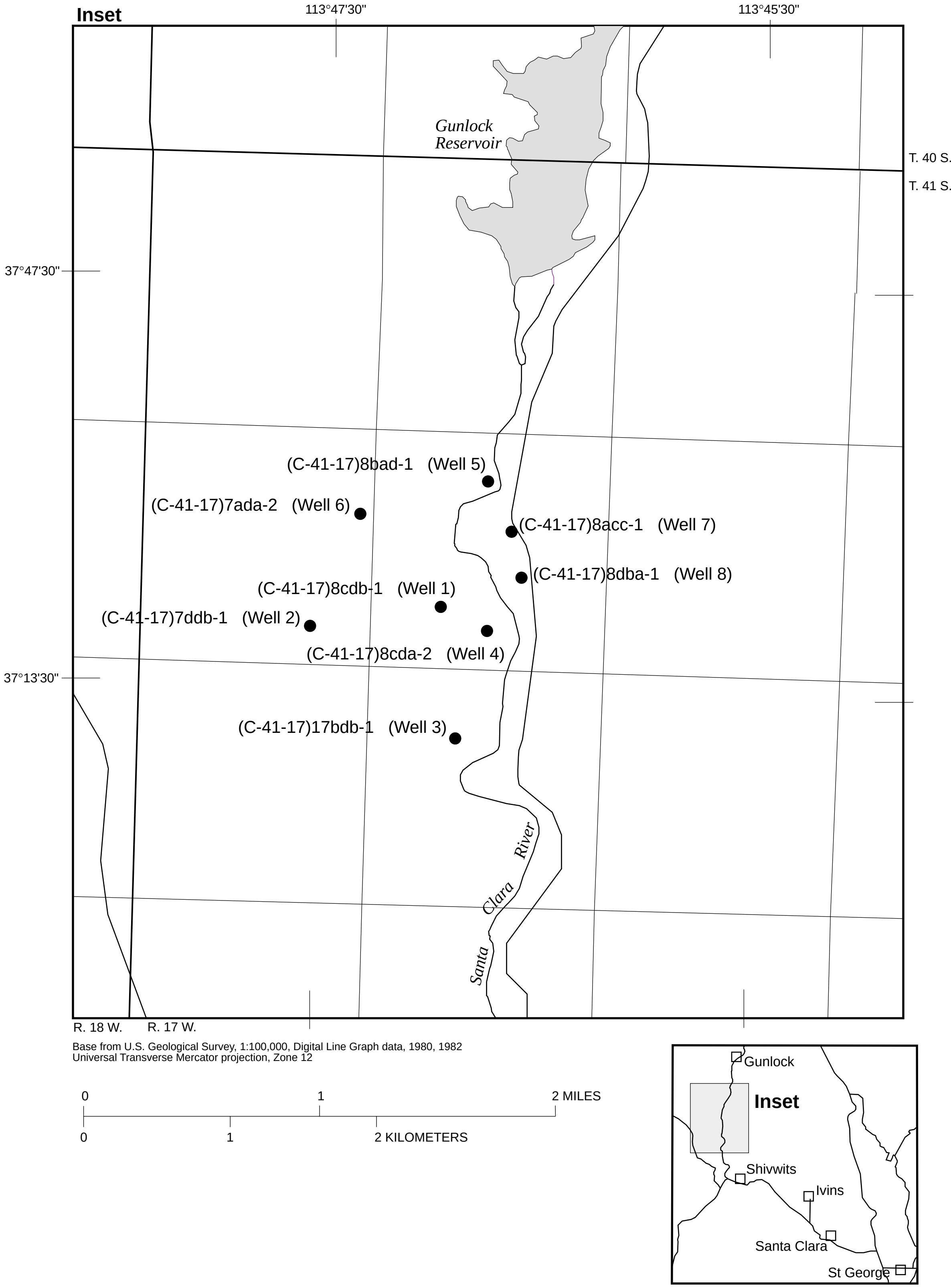


Figure A-9. Location of wells in the Gunlock aquifer test, Washington County, Utah, February 1996.

the test, during the 6 days of pumping, and for about 7 days after the pump was shut off. The pumped well and all of the observation wells are finished in the Navajo aquifer. Data for the pumped well and observation wells is reported in table A-3. Five of the observation wells (Gunlock wells 1, 4, 5, 6, 8) are production wells that had not been pumped for at least 19 days prior to the aquifer test. The two farthest observation wells are production wells that maintained a constant pumping rate both before and during the aquifer test. Gunlock well 3 (radial distance = 4,400 ft) was pumping at about 840 gal/min and Gunlock well 2 (radial distance = 4,855 ft) was pumping at about 570 gal/min.

Geology

At the aquifer-test site south of Gunlock Reservoir, the Navajo Sandstone is exposed at the surface and, because of erosion, is about 1,100 ft thick. As the Navajo Sandstone dips to the north-northeast, its thickness increases to a maximum of 3,000 ft at the contact with the overlying Carmel Formation about 1.5 mi north of the pumped well (Gunlock well 7). The Navajo Sandstone thins toward the southwest as a result of erosion until the geologic contact with the underlying Kayenta Formation is exposed about 2 mi southwest of the pumped well. The Navajo Sandstone is continuous for about 6 mi toward the northwest, beyond which it is offset completely by faulting. Similarly, the Navajo Sandstone is completely offset by the Gunlock Fault about 1 mi to the east of the pumped well. A generalized geologic cross section in the vicinity of the pumped well is shown in figure A-10.

Surface-fracture studies of outcrop sites near the pumped well and lineament studies of areal photographs indicate that the sandstone is highly fractured in this region (Hurlow, 1998). Rose diagrams of these fracture and lineament orientations indicate that the principal direction of fracturing ranges from due north to northwest. Field observations show a predominant fracture trend in the due north-south direction. The Santa Clara River follows this fracture trend from just downstream from the Gunlock Reservoir to a bend in the river by Gunlock well 5. The river then bends to the west until it turns south again and follows another parallel fracture set to a location adjacent to the pumped well (Gunlock well 7). These north-south fracture sets were observed to be much more continuous and have wider apertures compared to other fractures exposed along the outcrop. On the basis of this surface fracturing, it is assumed that the aquifer is anisotropic and hydraulic conductivity is higher in this direction.

In addition to the Navajo Sandstone, fluvial unconsolidated deposits are along the Santa Clara River valley. The width of these fluvial sediments generally is less than a few hundred feet at the aquifer-test site. The depth of these sediments is unknown.

Hydrology

The Santa Clara River flows within 600 ft of the pumped well. The amount of water in the river along the reach near the pumped well depends on the

Table A-3. Construction data for wells used in the Gunlock aquifer test, Washington County, Utah, February 1996

Gunlock well number	Well number	Radial distance (feet)	Casing diameter (inches to feet)	Open interval (feet below land surface)	Opening type
7	(C-41-17)8acc-1	0	16 to 800	200 - 800	Screen
8	(C-41-17)8dba-1	710	16 to 800	200 - 800	Screen
5	(C-41-17)8bad-1	1,650	16 to 384	100 - 384	Perforations
4	(C-41-17)8cda-2	2,000	16 to 573	123 - 573	Screen
1	(C-41-17)8cdb-1	2,100	16 to 283	100 - 200	Perforations
6	(C-41-17)7ada-2	3,530	16 to 573	123 - 573	Screen
3	(C-41-17)17bdb-1	4,400	16 to 9	9 - 626	Open hole
2	(C-41-17)16bbd-1	4,850	16 to 288; 10 to 466	176 - 466	Perforations

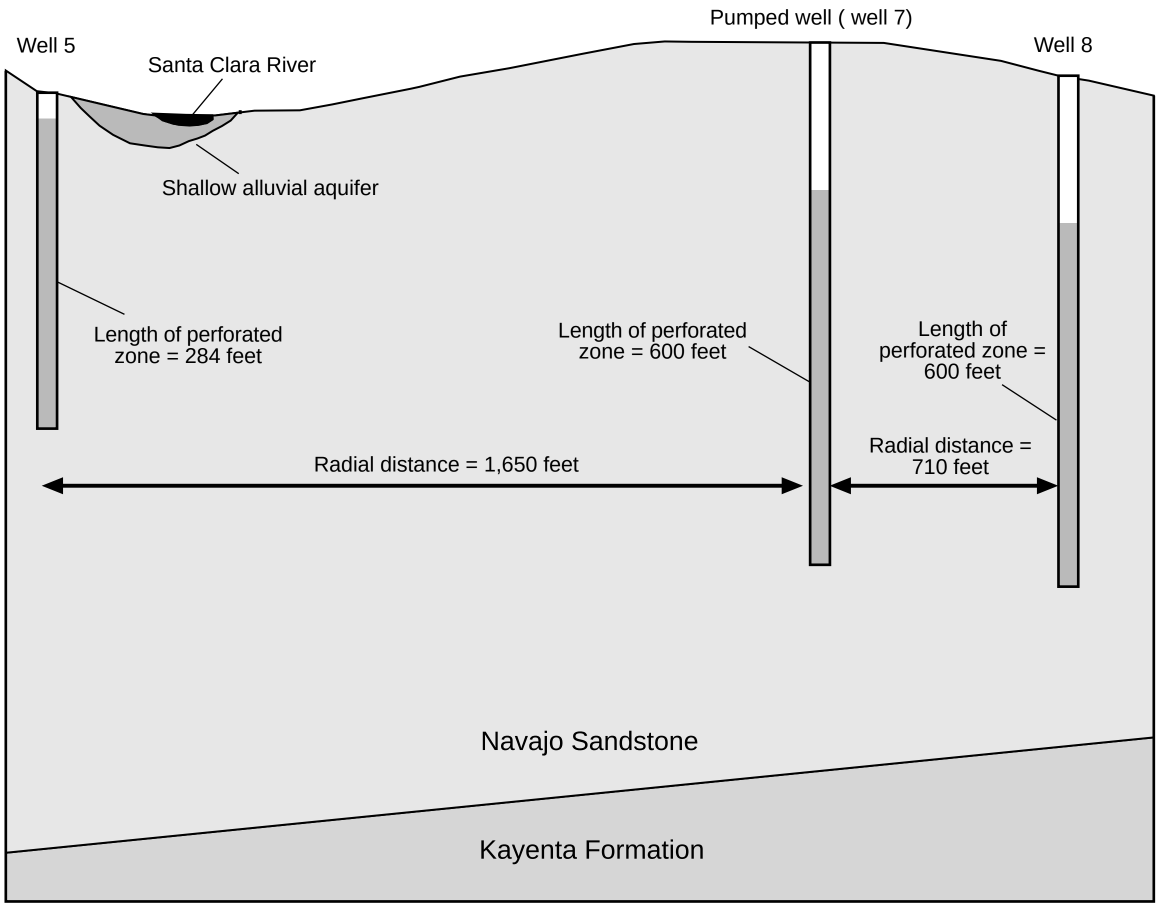


Figure A-10. Generalized geologic cross section in the vicinity of the pumped well in the Gunlock aquifer test, Washington County, Utah, February 1996.

upstream releases from Gunlock Reservoir. The valve controlling reservoir releases was closed more than a month before the aquifer test and was not opened until completion of the recovery part of the test. However, about 0.8 ft^3/s was leaking from the base of the reservoir before and throughout the aquifer test. Flow in the river gradually decreased southward to a point about 4,000 ft south of the pumped well where the river bed was dry prior to the start of the aquifer test.

Prior to the aquifer test, a staff gauge was installed in the river adjacent to well 5 (about 1,500 ft north of the pumped well), a 6-in. Parshall flume was installed in the river adjacent to well 7, and a 3-in. Parshall flume was installed in the river south of well 8 (fig. A-9). Staff-gauge measurements adjacent to well 5 indicate that flow upstream from the pumped well was constant during both the pumping and recovery parts of the aquifer test. However, discharge measurements at both flumes indicated that a minimum of about 110 gal/min (0.24 ft^3/s) was induced from the river into the shallow fluvial aquifer by the decrease in head in the underlying Navajo aquifer during the pumping part of the aquifer test. Because decreases in discharge downstream from the lower flume could not be measured (but are assumed to have occurred, as evidenced by the drying up of that river reach), the total amount of water lost from the river as a result of pumping was probably larger.

Although no observation wells are located in the shallow fluvial aquifer, head decreases in the Navajo aquifer caused by pumping were assumed to induce additional water from the fluvial aquifer into the Navajo aquifer. For a hypothetical calculation, the following assumptions were made: (1) the average thickness of the fluvial aquifer is 20 ft; (2) the average width of the fluvial aquifer is 100 ft; (3) the effective porosity of the fluvial sediments is 20 percent; and (4) head in the fluvial aquifer decreased an average of 0.5 ft along the 5,500-ft reach, which showed a decrease in discharge during the aquifer test. The volume of water released by this 0.5 ft drop in water level in the fluvial aquifer would be about 8 million gal—about the same total volume of water pumped during the entire aquifer test. Although some part of the pumped water came from storage within the Navajo aquifer, most of the water moving toward the pumped well during the aquifer test was assumed to be induced flow from the shallow fluvial aquifer and the Santa Clara River.

The saturated thickness of the Navajo aquifer is estimated to range from about 600 ft at well 3 to about 1,100 ft at well 5. The saturated thickness at well 7 (the pumped well) is about 1,050 ft when not pumped. After pumping equilibrium has been established, the saturated thickness decreases to about 800 ft. The pumped well is perforated for a 600-ft interval during static conditions and for a 550-ft interval during pumping conditions. Therefore, the perforated interval during pumping at well 7 is more than $^2/_3$ of the total saturated thickness at the well site. The observation wells are generally perforated in the same upper part of the Navajo aquifer. The closest observation well (Gunlock well 8, radial distance = 710 ft, total drawdown of 21.5 ft) has a nearly identical perforated interval. The other observation well that had substantial drawdown (Gunlock well 5, radial distance = 1,650 ft, total drawdown of 1.2 ft) is perforated in the uppermost 280 ft of the aquifer (fig. A-10). However, because its radial distance is three times the vertical perforated interval (550 ft during pumping) of the pumped well, partial penetration effects should be negligible.

Data Reduction and Analysis

Measured water levels at the observation wells were not corrected for barometric changes. The magnitude of drawdown and recovery at the nearest observation well (well 8) was much larger than effects resulting from barometric changes (generally less than 1 ft). A comparison between barometric pressure and prepumping water levels at Gunlock well 5 did not show any correlation. Therefore, no corrections for barometric pressure variations were attempted at this well and the more distant observation wells (Gunlock wells 1, 4, and 6). Water-level increases of from 7 to 9 ft were measured at Gunlock wells 1, 4, and 6 throughout the prepumping and recovery parts of the aquifer test. After linear trend corrections were applied to the water-level data from these wells, the recovery data indicate that these wells were only affected slightly by pumping at well 7 (about 0.3 ft at each well), not enough to produce drawdown curves of sufficient quality for curve fitting.

Because of small variations in the pumping rate throughout the pumping part of the aquifer test, the observation-well recovery data were used for wells 5 and 8. The only corrections made to the recovery data for these two wells were to subtract the prerecovery trend. To determine the prerecovery trend at wells 5 and 8, a straight line was fitted to the latter part of a semilog plot of the prerecovery data. This trend was then extended for the recovery part of the test and added to uncorrected recovery. The drawdown and recovery data for these two observation wells were then plotted out together on a log-log scale by dividing the elapsed time by the observation well's radial distance squared.

Initial attempts to match the observed recovery curves for the two wells with the Theis solution (1935) for confined aquifers did not provide a satisfactory match; the Theis curve matches early time recovery data from well, but then deviates at later time (fig. A-11). The later-time observed drawdown is less than predicted by the type curve, which may indicate additional sources of water besides release of water from confined storage. No confining layer is present at the aquifer-test site, but early time responses at more distant observation wells initially appear to reflect confined conditions. Therefore, curve-fitting with the Neuman (1974) unconfined solution with delayed yield was attempted. Although the delayed-yield curve (the lower of curve in fig. A-12) provided a better individual match to the data from well 5, a single simultaneous solution for transmissivity and storage was not possible for both wells. The modified Hantush solution (Lohman, 1972, p. 32-34) for leaky confined aquifers also was attempted with the assumption that leakage from the overlying fluvial sediments would be similar to an overlying leaky layer, but an acceptable single-value solution could not be achieved. It is assumed that the large difference between the two well data sets may be a result, in part, of anisotropic conditions. Homogeneous and anisotropic conditions are indicated if the recovery data sets have offset but parallel late-time slopes, as shown earlier by applying a modified form of the Papadopulos solution (1965) to the Anderson Junction aquifer test. However, later-time data on a semilog plot of recovery from the two Gunlock observation wells do not have similar slopes (fig. A-13). Therefore, the response at the two observation wells is assumed to be a combination of (1) anisotropic conditions that resulted from fracturing, and (2) leakage from the overlying river and fluvial aquifer (a partially penetrating boundary). There is no analytical method that can be used for this complex hydrologic setting.

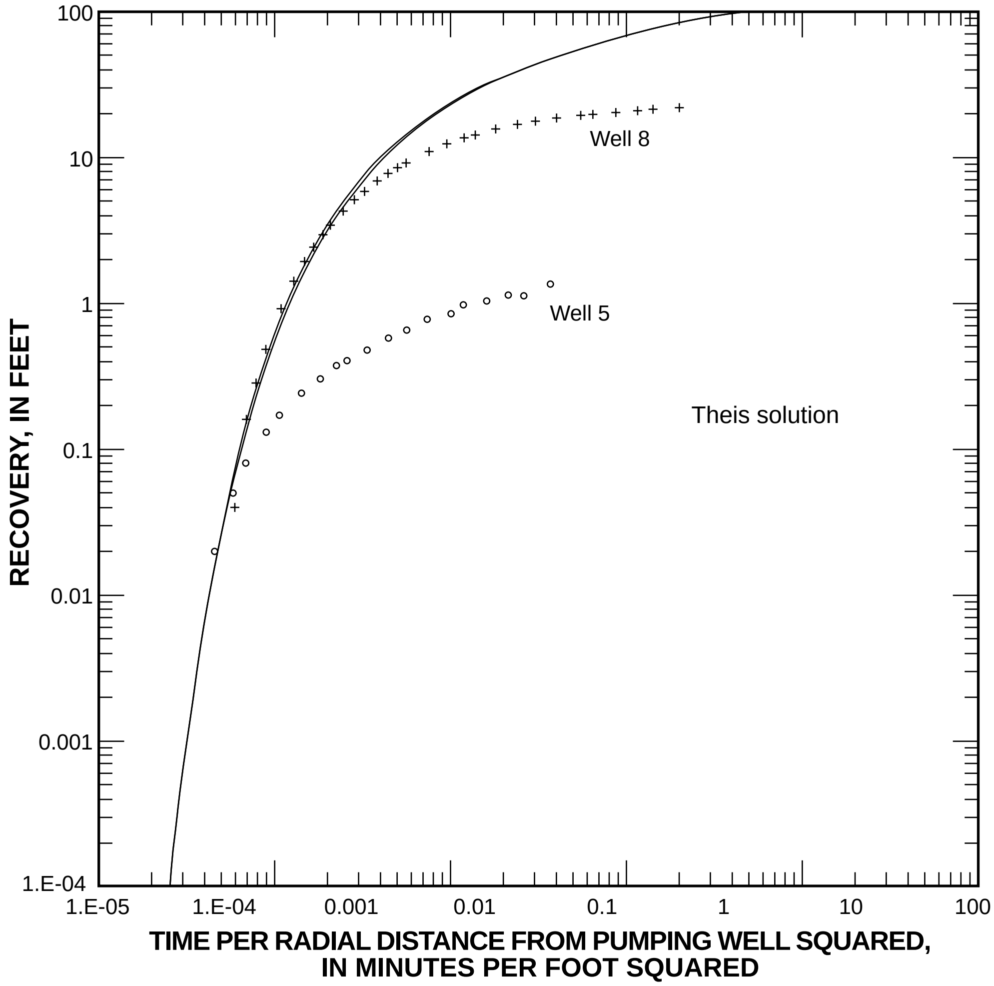


Figure A-11. Recovery data from two observation wells during the Gunlock aquifer test, Washington County, Utah, February 1996 (Theis solution, 1935).

Ground-Water Flow Model

To analyze the observation-well data from the Gunlock aquifer test, a three-dimensional ground-water flow model was constructed and calibrated using Modflow 96 (Harbaugh and McDonald, 1996). The ground-water flow model was developed as a tool for aquifer-test analysis and therefore uses the principles of superposition to simulate the change in heads and flows that resulted from pumping at well 7. As stated by Reilly and others (1987, p. 2), “The principle of superposition means that for linear systems, the solution to a problem involving multiple stresses is equal to the sum of the solutions to a set of simpler individual problems that form the composite problem.” In general, the principal of superposition can only be applied to a confined aquifer. However, Reilly stated that the principle of superposition can be applied to mildly nonlinear systems such as an unconfined aquifer if the regional drawdown that results from pumping is less than 10 percent of the full saturated thickness of the aquifer. This is the case at the Gunlock aquifer-test site. The regional dewatering of the aquifer by pumping from well 7 represented only a very small percentage of the prepumping saturated thickness. By using the principle of superposition, only the changes in simulated heads and flows from pumping need to be analyzed. To isolate these changes, absolute elevation data were converted to relative elevation data such that prior to pumping, the water table everywhere in the model was at 0 ft. The initial conditions, rather than being specified in absolute terms (actual head values in ft above sea level), are specified relative to the heads and flows that existed prior to pumping at well 7.

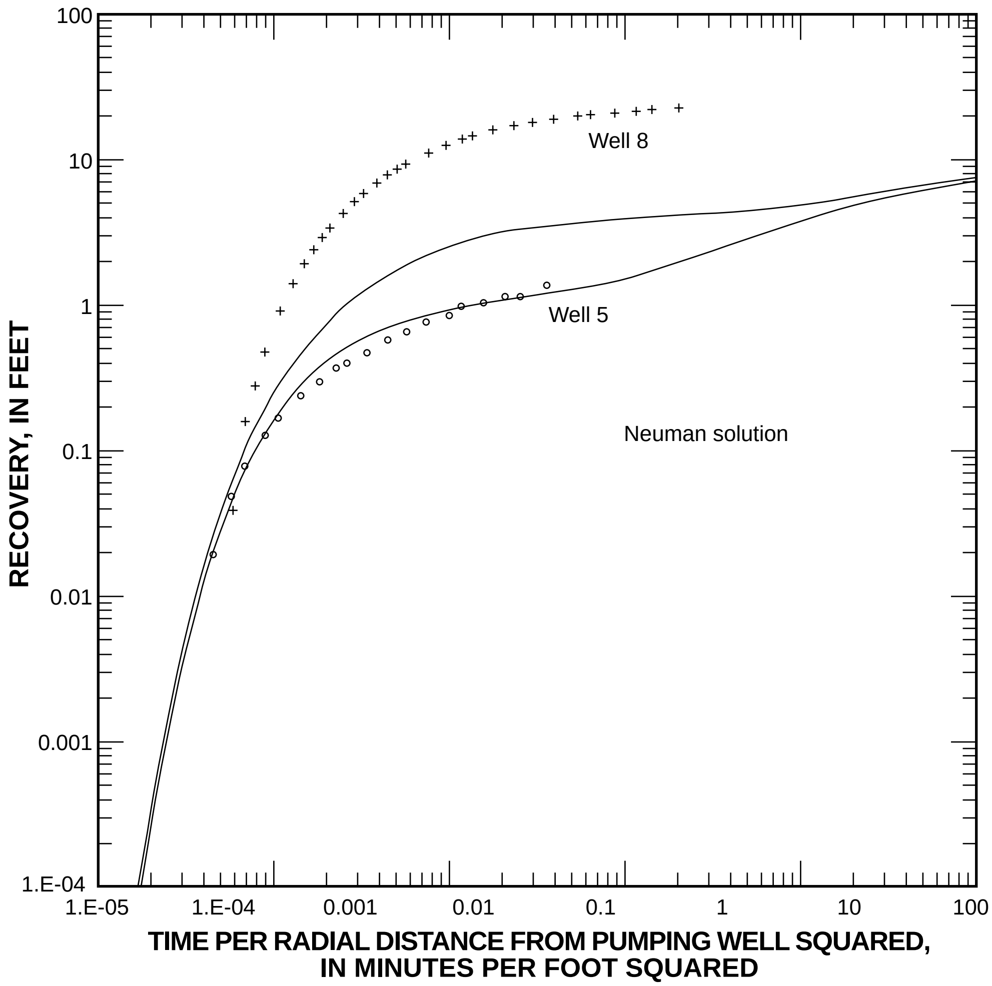


Figure A-12. Recovery data from wells during the Gunlock aquifer test, Washington County, Utah, February 1996 (Neuman solution, 1974).

The location of the model boundary with respect to the Gunlock part of the Navajo aquifer is shown in figure A-14a. The model was discretized into 163 rows by 149 columns. The cell size at the center of the model is about 10 ft by 10 ft (fig. A-14c)and increases with radial distance from the pumped well to a maximum cell size of about 400 ft by 400 ft along the perimeter of the model using a multiplier of approximately 1.5 (fig. A-14b). The active area of the model is surrounded by a no-flow boundary. The base of the model (bottom of layer 1) is also a no-flow boundary because published hydraulic-conductivity values for the Kayenta Formation determined from laboratory analyses are generally lower than values for the Navajo Sandstone (Weigel, 1987).

Because the model has only one layer that represents the Navajo aquifer, the combined effects of seepage from the river and shallow fluvial aquifer were simulated by using the River Package. The fluvial aquifer is not simulated as a separate layer because there is no available data on its geometry, aquifer properties, or water levels. Conductance values of river cells were varied during model calibration to match measured losses along the Santa Clara River. The stage of the river was specified at 0 ft everywhere, the same elevation as the top of the aquifer and the defined initial head value. Thus, until the stress from pumping propagated out to the nearest river cells, no seepage from the river would be simulated. In this manner, the changes in stream seepage rates as a result of pumping at well 7 could be isolated and evaluated.

The Well Package is used to simulate pumping at well 7. The specified pumping rate was 845 gal/min for the stress period representing the pumping part of the aquifer test. Because Gunlock Wells 2 and 3 were also pumping both before and throughout the aquifer test,

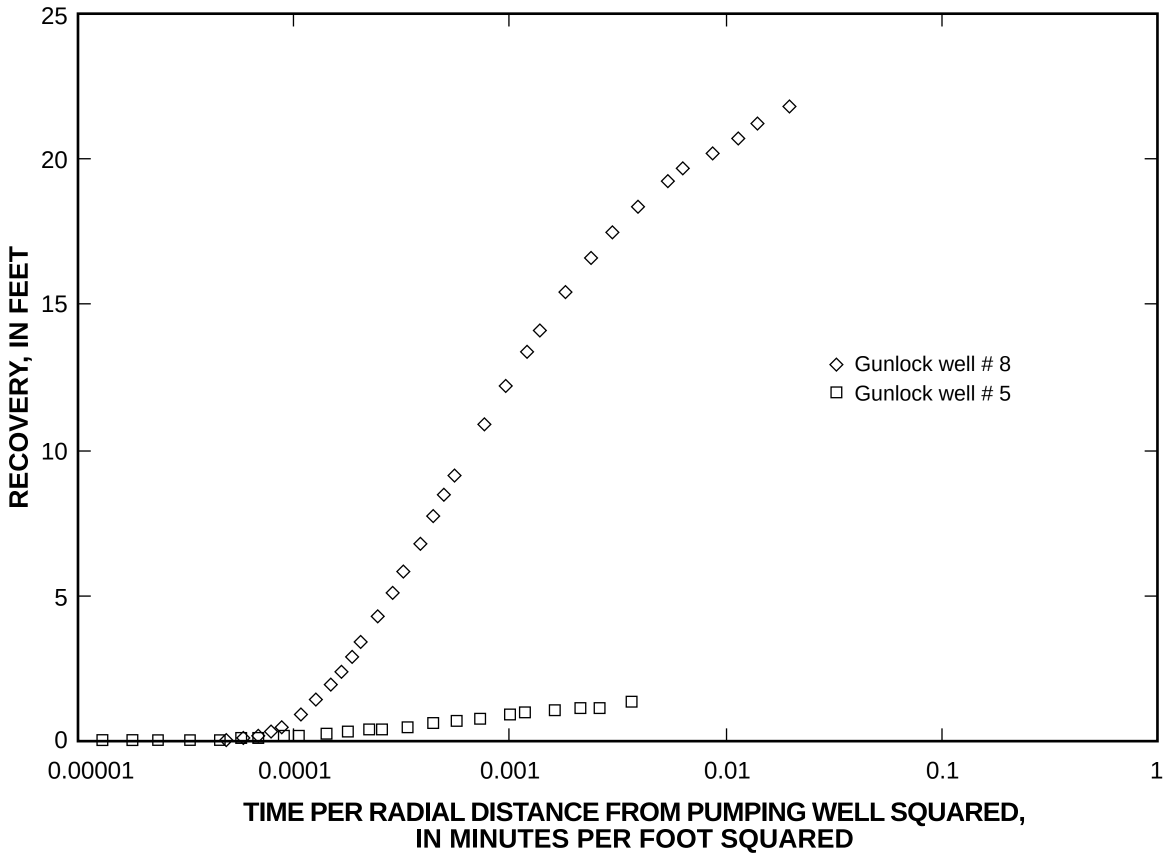


Figure A-13. Recovery data showing different late-time slopes for wells during the Gunlock aquifer test, Washington County, Utah, February 1996.

these were not simulated as additional stresses; the model was constructed to evaluate only changes resulting from pumping at well 7.

Model Calibration

The parameters used to calibrate the ground-water flow model are (1) drawdown curves at the nearest two observation wells (wells 5 and 8), (2) total drawdown at well 7 and at more-distant observation wells, (3) ground-water budget parameters, (4) anisotropy resulting from fracturing, and (5) known aquifer boundaries.

Matching measured drawdown/recovery curves at wells 5 and 8 was the most important calibration point of the model. The final match of computed drawdown to measured recovery is shown in figure A-15. Generally, the computed drawdown matches the measured drawdown at both observation wells at early and late time. At “middle” time, the computed drawdown values are slightly less than measured values. The lack of a perfect match is probably because of the simplifying assumptions, such as homogeneity in aquifer properties, uniform anisotropy in the north-south direction, and the assumption of horizontal flow in a single-layer model.

Matching total drawdown at the pumped well and distant observation wells was not as high a priority as matching drawdown curves at the nearby observation wells. Nevertheless, this was considered important information for the calibration. The total computed drawdown of 303 ft at the pumped well was more than the 257 ft of measured drawdown. However, matching drawdown at the pumped well is complicated by finite-difference limitations and the poor-quality data associated with pumped well measurements. Wells 1, 4, and 6 had similar computed-versus-measured total drawdown values. As mentioned earlier, because these wells were undergoing substantial recovery during the aquifer test, the corrected drawdown values computed from water-level measurements may contain some error. Well 3 displayed no measurable drawdown during the pumping part of the aquifer test. However, both this well and well 2 (outside the active model boundary) were pumping

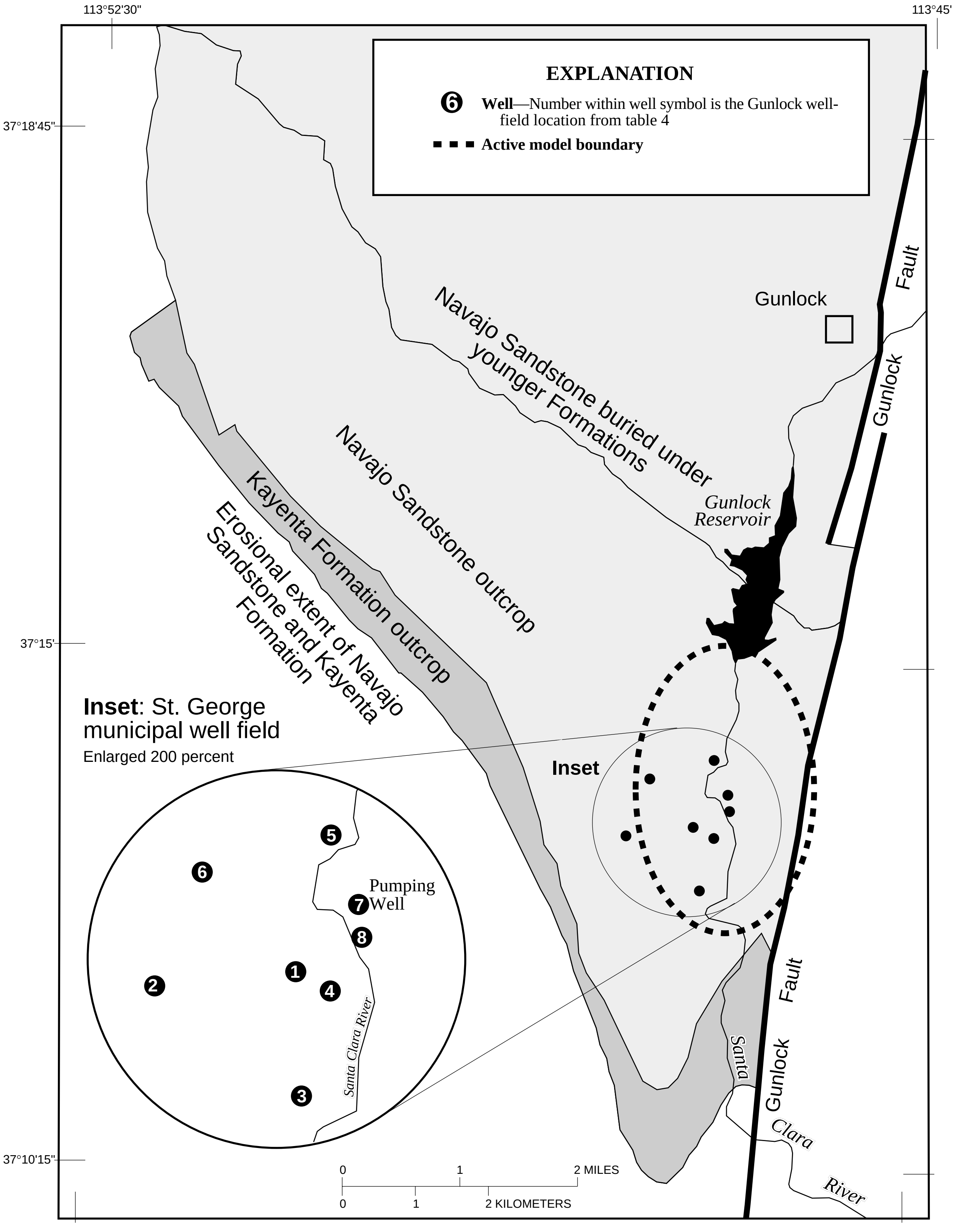


Figure A-14. (a) Boundary, (b) finite-difference grid, and (c) detail of finite-difference grid for the ground-water flow model of the Gunlock aquifer test, Washington County, Utah, February 1996.

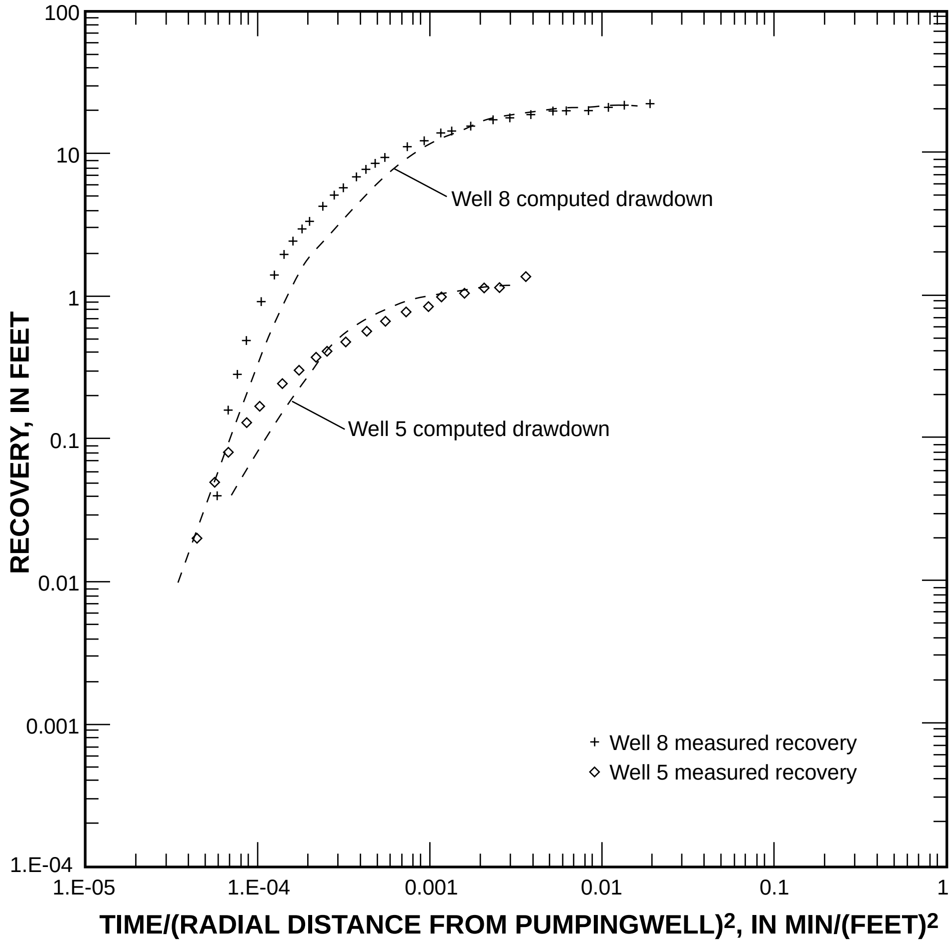


Figure A-15. Measured recovery and computed drawdown for wells during the Gunlock aquifer test, Washington County, Utah, February 1996.

during the test, so it was not possible to determine if there were very small effects at these wells.

In the ground-water flow model, more than 90 percent, or 760 gal/min, of water discharging from the aquifer at the pumped well came from the River Package (simulating both the Santa Clara River and shallow fluvial aquifer) at the end of the 6-day pumping period. This is much more than the measured 110 gal/min loss from the Santa Clara River but also includes the dewatering of the shallow fluvial sediments that was not measured during the aquifer test. The other 10 percent, or 80 gal/min, came from aquifer storage.

As discussed above, the predominant orientation of surface fracturing on the exposed Navajo Sandstone outcrop near the pumped well is north-south. Although the orientation of preferential flow as a result of fracturing was a constraint in developing the ground-water flow model, there was no prior information regarding the relative degree of anisotropy. Therefore, anisotropy factors for $K_{north\text{-}south}$: $K_{east\text{-}west}$ from 1:1 to 10:1 were tried during the calibration process. The final calibrated model uses an anisotropy factor for $K_{north\text{-}south}$: $K_{east\text{-}west}$ of 3:1.

As discussed earlier, the Gunlock part of the Navajo aquifer has a limited extent as a result of faulting and erosional boundaries to the east, south, and west. However, water-level measurements during the aquifer test indicated that the drawdown cone had not reached any of these boundaries. Similarly, after 6 days of simulated pumping, the ground-water flow model did not produce substantial drawdown at these boundaries. Simulated drawdown at the nearest boundary, the Gunlock Fault to the east, was less than 0.5 ft. It is possible, however, that long-term pumping at well 7 may result in noticeable boundary effects at the observation wells, such as increased rate of drawdown with time.

Generally, the model was more sensitive to changes in hydraulic conductivity and anisotropy ratios and less sensitive to changes in storage and riverbed conductance. However, any general statements regarding relative sensitivity may oversimplify a more complex situation. For example, although order-of-magnitude changes in storage may not affect total drawdown substantially at the pumped well and nearby observation wells, they strongly affect total drawdown and the shape of the drawdown cone at greater radial distances, as well as the water-budget components. Similarly, although order-of-magnitude changes in riverbed conductance may not cause substantial changes to the water-budget components and the extent of the drawdown cone, such changes strongly affect drawdown at observation wells.

There are two important limitations to the calibrated ground-water flow model and its use as a tool for analysis of aquifer-test data from the Gunlock site. First, a single-layer model does not simulate flow in the shallow fluvial sediments along the Santa Clara River, nor allow for the simulation of vertical ground-water flow and determination of vertical anisotropy. If another aquifer test is to be conducted at this site, it would be helpful to drill a few shallow observation wells into the fluvial aquifer to determine hydrologic properties of these sediments, thickness of the fluvial aquifer, and drawdown caused by pumping from the Navajo aquifer. These data could be used to construct an additional model layer representing the shallow fluvial aquifer. Second, anisotropic conditions are assumed to be consistent throughout the modeled area. Differences in fracture density and orientation at the aquifer-test site may result in a varying degrees of anisotropy. Because detailed data about the variation in fracturing both laterally and vertically are not available for the site, aquifer properties were assumed to be uniform throughout the simulated area. Additional surface- and borehole-fracture data at the site may help to identify the variability in anisotropy due to fracturing.

Summary

The values determined from model calibration are 0.33 ft/d for horizontal hydraulic conductivity in the east-west orientation and 1.0 ft/d for horizontal hydraulic conductivity in the north-south orientation. Multiplying these values by the aquifer thickness of 1,100 ft at the pumped well results in transmissivity values of about 360 to 1,100 ft^2/d.

The range of hydraulic-conductivity values determined from this aquifer test are similar to Cordova's (1978, p. 26) laboratory determination of horizontal hydraulic-conductivity values that ranged from 0.36 to 5.0 ft/d for samples from the Navajo aquifer at various locations within Washington County. They are also similar to the horizontal hydraulic-conductivity value of 0.8 ft/d determined from the Hurricane Bench aquifer test. However, the values are lower than the range of horizontal hydraulic-conductivity values of 1.3 to 32 ft/d determined from the Anderson Junction aquifer test. Because the Navajo Sandstone is composed of well-sorted very fine sand and varies little throughout southwestern Utah, the higher values of horizontal hydraulic conductivity determined from the Anderson Junction aquifer test are probably because of a higher degree of fracturing (higher fracture density and larger average aperture).

The value for storage coefficient determined from model calibration is 0.001. This value is the same order-of-magnitude as the value of 0.002 determined from the Hurricane Bench aquifer test and the value of 0.0013 determined from the Anderson Junction aquifer test.

Grapevine Pass Aquifer Test

The purpose of the Grapevine Pass aquifer test was to determine the transmissivity of the Navajo aquifer near Grapevine Pass, about 7 mi northeast of St. George in Washington County, Utah (fig. A-16). The aquifer test was conducted in February 1996 by the USGS in coordination with the Water Department of Washington, Utah. Unlike the other aquifer tests, this was a single-well aquifer test with drawdown and recovery measured only in the pumped well. Water levels were measured in well (C-41-15)28dcb-2 during the 24 hours prior to the test, during the 24 hours of pumping, and during the 24 hours after the pump was shut off. Water from the pumped well was diverted into nearby Grapevine Pass Wash and removed from the aquifer-test site. Discharge was estimated to average 180 gal/min and was measured with both a v-notch weir and a pito tube attached to the discharge pipe.

According to field observations, geologic maps, and surface-fracture surveys, the Navajo Sandstone outcrop in the immediate vicinity of the aquifer test site has no prominent surface fracturing (Hurlow, 1998). It was noted, however, that surface fractures are present within about 1 mi of the site, both up- and down-canyon. According to the drillers' log, the Navajo Sandstone is about 915 ft thick at the site and is interbedded with layers of siltstone and mudstone. The drillers' log

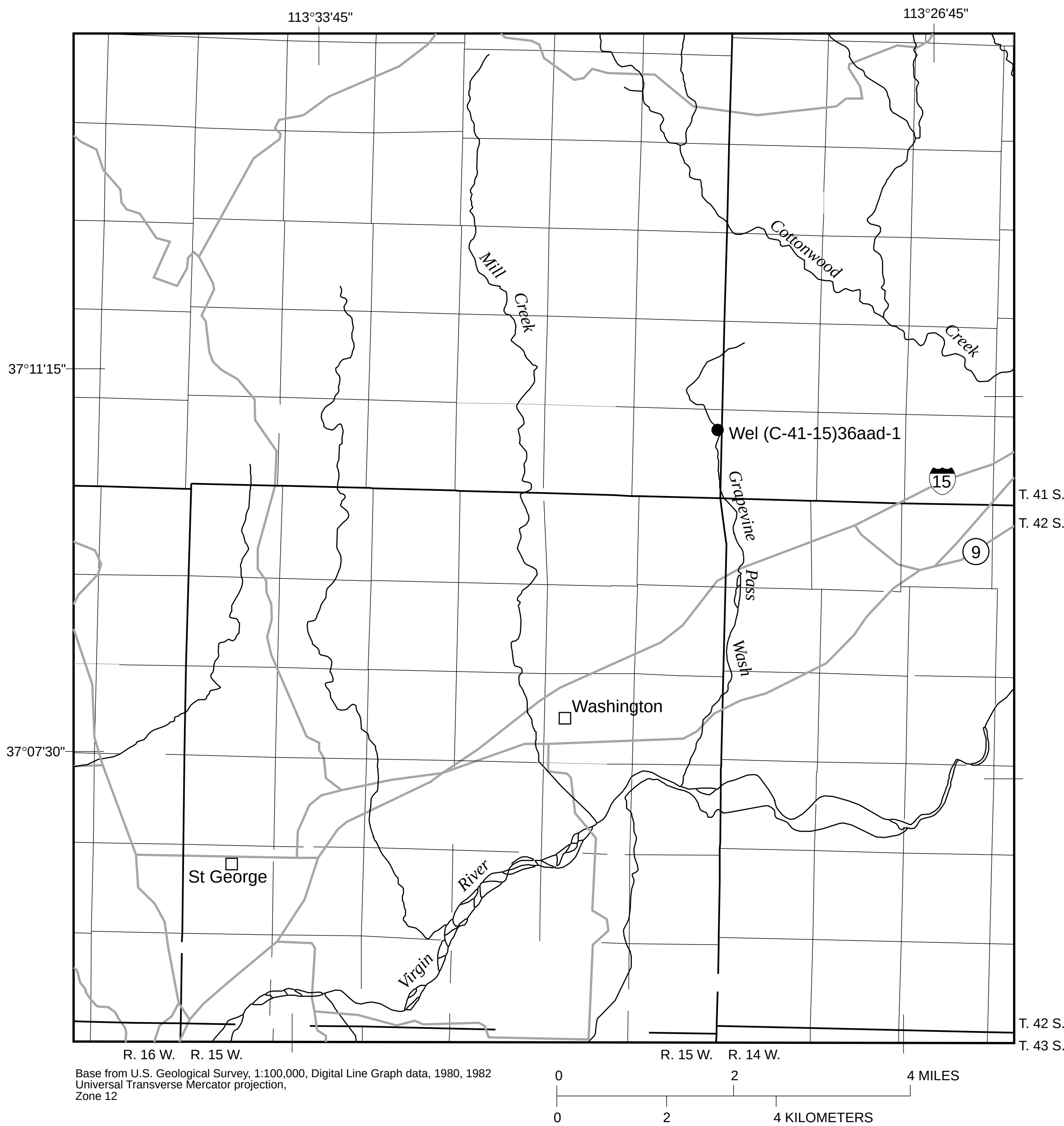


Figure A-16. Location of well in the Grapevine Pass aquifer test, Washington County, Utah, February 1996.

notes that the Navajo Sandstone has a much smaller grain size there than at other wells completed in the Navajo Sandstone. The Kayenta Formation, made up of siltstone with intermixed clays and sands, is present from a depth of 920 ft to the bottom of the drillhole at 950 ft. The Utah Geological Survey also analyzed borehole cuttings from three wells drilled in the Navajo Sandstone in Washington County: the Grapevine Pass production well, the Anderson Junction production well about 20 mi northeast of St. George, and a production well in the Winchester Hills subdivision about 7 mi north of St. George. When compared with lithologic analyses from the two other wells, the Grapevine Pass site had much more interbedding with finer siltstone and mudstone layers (J. Wallace, Utah Geological Survey, written commun., 1996). Therefore, the fine-grained material at this site and the lack of surface fracturing may indicate lower hydraulic conductivity in this area. After the well was completed, the static water level was about 350 ft below land surface, which indicated a saturated thickness of about 570 ft for the Navajo aquifer.

The Cooper-Jacob straight-line method was chosen for analysis of the data. The semilog plot of recovery versus time used for the analysis is shown in figure A-17. The early time recovery data apparently are affected by well-bore storage effects, as a result of a combination of the large diameter well casing (12 in.) and the very small perforations necessary to keep the fine grained sand matrix of the aquifer from entering the casing. A method outlined in “Groundwater and Wells” (Driscoll, 1986, p. 232 -235) shows an interpretive technique for determining the critical time when the borehole-storage effect becomes negligible when using the following equation (eq. 9.9, p. 233):

$$t_c = \frac{0.6([d_c]^2 - [d_p]^2)}{Q/s} \qquad \text{(A16)}$$

where

t_c is the time in minutes when casing storage becomes negligible,

d_c is the inside diameter of the well casing in in.,

d_p is the outside diameter of the pump column pipe in in., and

Q/s is the specific capacity of the well in gal/min/ft of drawdown at t_c.

For the Grapevine Pass aquifer test, d_c = 12 in.; d_p = 4.23 in.; Q = 180 gal/min. Assuming an initial recovery *(s)* of 200 ft, the estimated initial iteration is:

$$t_c = \frac{0.6([12]^2 - [4.23^2])}{180/200} = 84 minutes \qquad \text{(A17)}$$

From the semilog recovery plot, at t = 84 minutes, the recovery is 362 ft. Solving for t_c with a recovery value of 362 ft yields t_c = 152 minutes for the second iteration. Working through this process for two more iterations yields a value for t_c of 158 minutes. This value correctly estimates the break in slope shown in figure A-17.

Thus, the Cooper-Jacob straight-line method (Cooper and Jacob, 1946) is used for time greater than 158 minutes. However, fitting a straight line to the recovery data beyond 158 minutes does not yield one unique fit. Two possible matches (lines T_1 and T_2) are shown in fig. A-17. The calculated transmissivity values from these lines are 160 ft^2/d and 330 ft^2/d, respectively. Because of this possible range of interpreted values, an order-of-magnitude value of 100 ft^2/d will be reported for this aquifer test. Assuming the maximum possible saturated aquifer thickness of about 500 ft, the horizontal hydraulic conductivity is about 0.2 ft/d. This is about one order of magnitude less than horizontal hydraulic-conductivity values determined from laboratory analysis of outcrop samples (Cordova, 1978, p. 26) and from the results of the other multiple-well aquifer tests. This lower hydraulic conductivity is consistent with the presence of finer-grained material and the lack of surface fracturing at this location.

New Harmony Aquifer Test

The purpose of the New Harmony aquifer test was to determine the transmissivity and storage properties of the Tertiary Pine Valley quartz monzonite along Ash Creek near New Harmony in Washington County, Utah (fig. A-18). The aquifer test was conducted during October and November 1996 by the USGS in coordination with the Church of Latter Day Saints Property Division. The multiple-well aquifer test involved pumping at well (C-38-13)35aba-1 for 7 days. The discharge from the pumped well was diverted into a 12-in. diameter pipe that carried the water to sprinkler pivots more than 1 mi away. Discharge throughout the test was estimated to average 1,050 gal/min (2.34 ft^3/s). Cumulative discharge was measured with an in-line flow meter. The average discharge rate was calculated by dividing the total number of gallons pumped by 7. Instantaneous discharge measurements were made with the in-line flow meter and a Clampatron meter through-

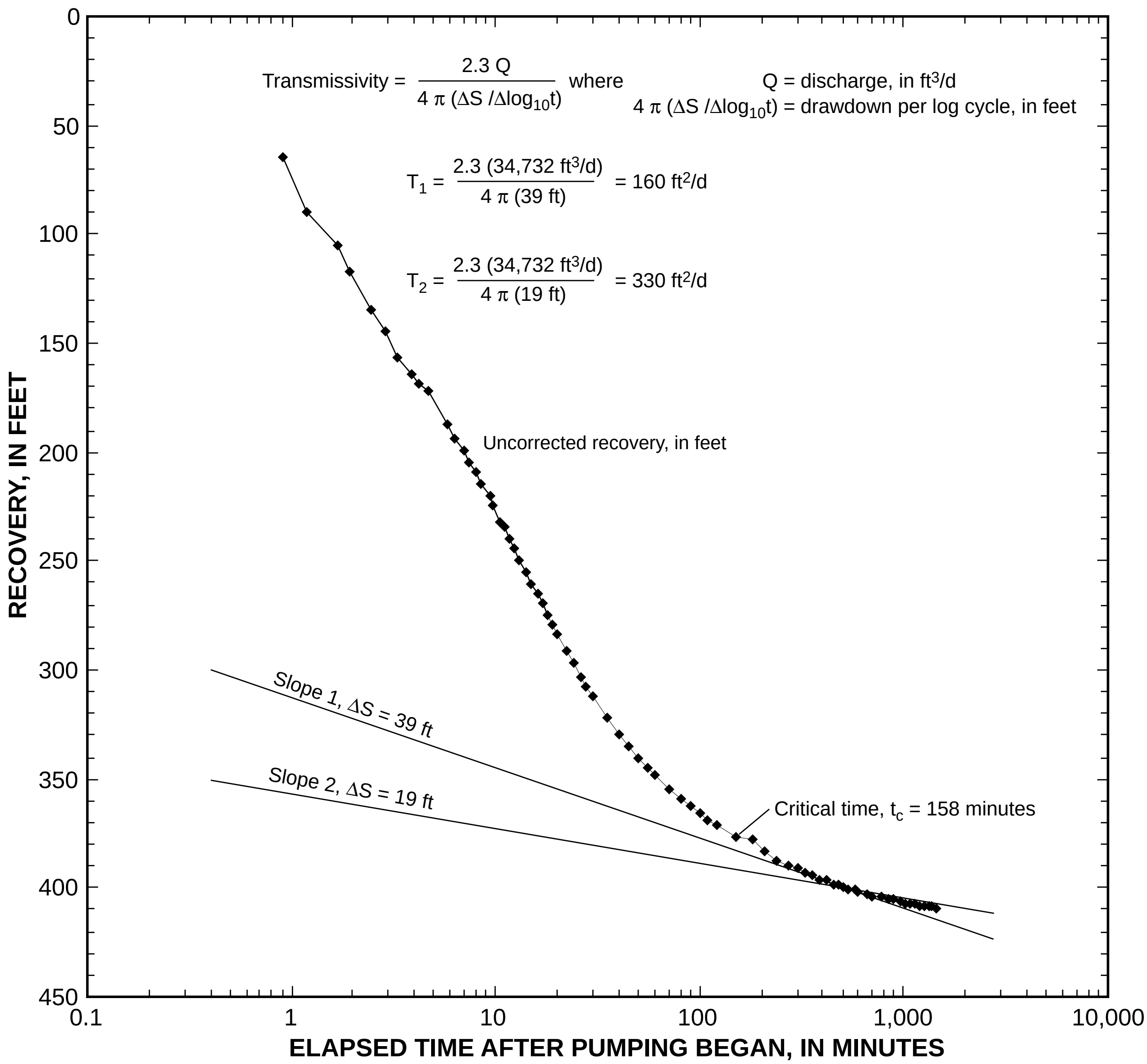


Figure A-17. Recovery data for the Grapevine Pass aquifer test using the Cooper-Jacob straight-line method, Washington County, Utah, February 1996 (Lohman, 1972).

out the test to ensure that the pumping rate did not fluctuate by more than 10 percent.

Water levels were measured manually in 10 observation wells and the pumping well beginning 18 days prior to the test, during 7 days of pumping, and for 7 days after the pump was shut off. Radial distances of the observation wells ranged from 825 to 7,950 ft. Data for the pumping well and observation wells are reported in table A-4. Observation well (C-38-13)35abb-1, referred to as the recorder well, was equipped with an automatic data recorder that continuously measured water levels beginning 18 days prior to the test, during the pumping part of the test, and for as much as 2 months after the pump was shut off. Because of the pumped well's proximity to Ash Creek, a flume was installed on the creek about 1 mi downstream of the well (and about 50 ft southwest of well (C-38-13)36cdd-1 to measure discharge. However, no decrease in flow was detected during pumping.

Hydrogeology

Based on drillers' logs and a geologic map by the Utah Geologic Survey (Hurlow, 1998), there is a 20- to 60-ft thick surficial layer of Quaternary fluvial material associated with Ash Creek at the aquifer-test site.

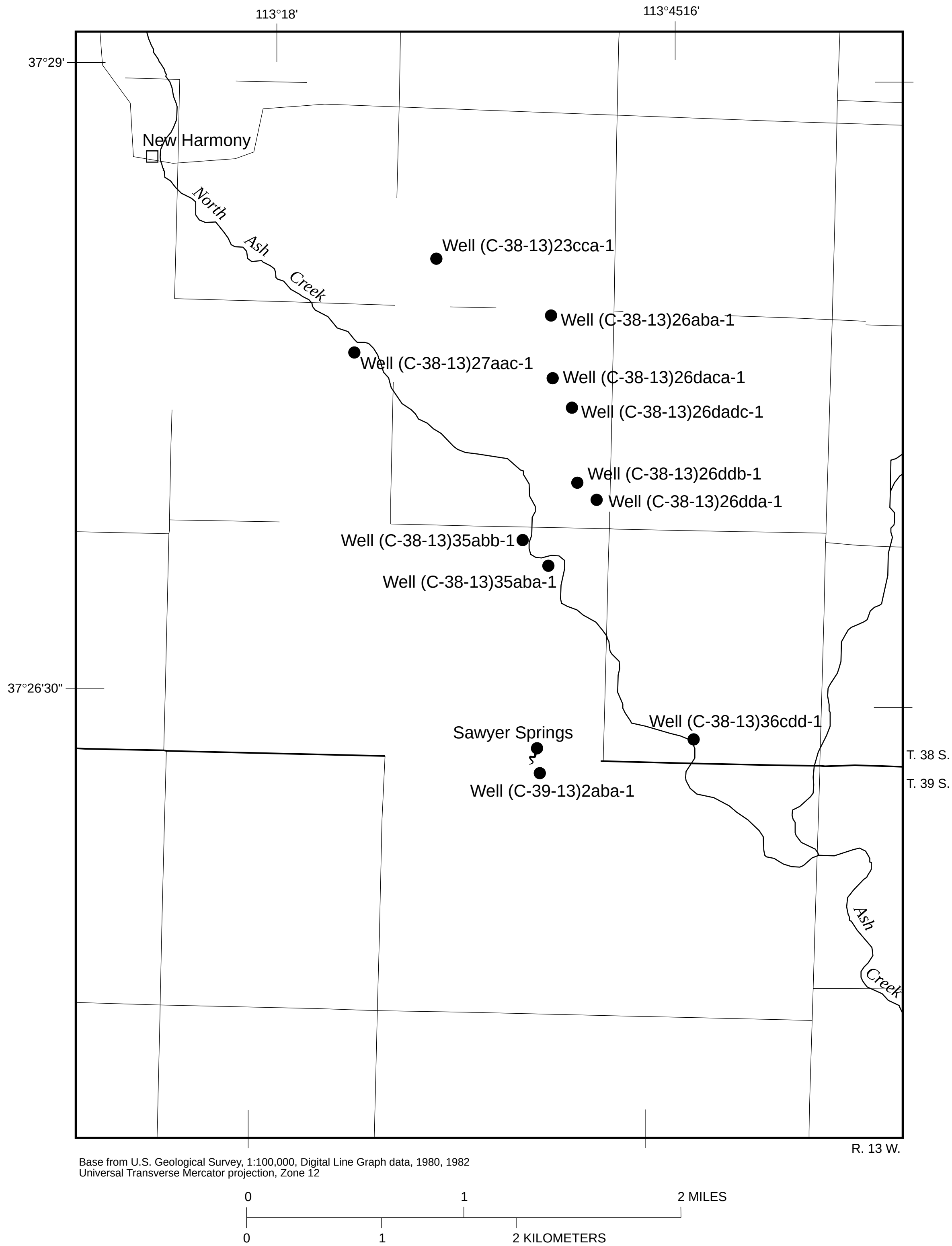


Figure A-18. Location of wells in the New Harmony aquifer test, Washington County, Utah, October and November 1996.

Table A-4. Construction data for wells observed during the New Harmony aquifer test, Washington County, Utah, October and November 1996
[NA, Not available

Well number	Radial distance (feet)	Casing diameter and length (inches to feet)	Open interval (feet below land surface)	Opening type	Geologic formation[1]
(C-38-13)35aba-1	0	12 to 620	220 - 620	Perforations	Tvip
(C-38-13)35abb-1	825	6 to 370	180 - 370	Perforations	Tvip
(C-38-13)26dda-1	2,100	6 to 200	160 - 200	Perforations	Tvip
(C-38-13)26ddb-1	2,150	6 to unknown	NA	NA	Tvip
(C-38-13)26adc-1	3,600	8 to 62; 6 to 199	40 - 199	Perforations	Tvip
(C-38-13)26aca-1	4,500	NA	NA	NA	NA
(C-38-13)36cdd-1	5,650	16 to 590	140 - 590	Perforations	Tvip
(C-39-13)2aba-1	5,800	16 to 400; 8 to 600	200 - 600	Perforations	Tvip
(C-38-13)26aba-1	6,050	6 to 177	150 - 175	Perforations	Tvip
(C-38-13)27aac-1	6,950	6 to 258	160 - 258	Perforations	Tvip
(C-38-13)23cca-1	7,900	12 to 130	36 - 122	Perforations	Qs

[1]See pl. 1 for definitions of geologic formations.

Underlying these unconsolidated sediments is the Tertiary Pine Valley quartz monzonite (Tvip), which is estimated to be as much as 3,000 ft thick. A schematic cross section through some of the wells at the aquifer-test site is shown in figure A-19. Although this fine-grained crystalline rock has low primary porosity, it has highly fractured zones capable of transmitting a large amount of ground water. Cook (1957, p. 73-75) describes fracturing in the Pine Valley quartz monzonite as follows:

"The basal 'dark brown zone' has a pseudocolumnar structure due to intersecting vertical joint sets and it forms vertical cliffs above the weak Claron limestone...The basal zone grades upward into the slightly less resistant but much thicker 'brown zone,' also greatly fractured by vertical joints...The purple zone rock has a pale reddish-purple groundmass and is much less jointed than the two lowermost zones. However, incipient jointing is often seen, marked by aligned, elongate weathering depressions."

A driller's log from the nearest observation well (C-38-13)35abb-1 indicates that there was a highly permeable fracture zone from 243 to 340 ft depth. Because there is no poorly permeable lithologic layer overlying the quartz monzonite, it is assumed that the aquifer is unconfined. It is possible, however, that areas with low fracture interconnectivity within the quartz monzonite may act as poorly permeable confining zones for underlying highly fractured zones.

Data Reduction and Analysis

During the aquifer test, only the recorder well (radial distance of 825 ft) showed substantial drawdown due to pumping. Measured water levels at this well were not corrected for barometric changes. The total change in water level at this well due to pumping was about 5.5 ft. The maximum possible change in water level due to fluctuating barometric pressure, assuming 100-percent efficiency, is only 0.2 ft, or 3.6 percent of the total change, and is therefore considered negligible.

The 5.5 ft of total drawdown at the recorder well was similar in order of magnitude to the 17 ft of total drawdown at the pumped well. The recorder well shares only 60 ft of the 400-ft open interval of the pumped well (fig. A-19); larger total drawdown at the recorder well would be expected if the open intervals of the two wells were the same. A plot of log-drawdown versus log-time data from the recorder well does not fit a Theis curve (fig. A-20). The Theis-curve solution shown in this figure was calculated with a storage value of 0.001 and a

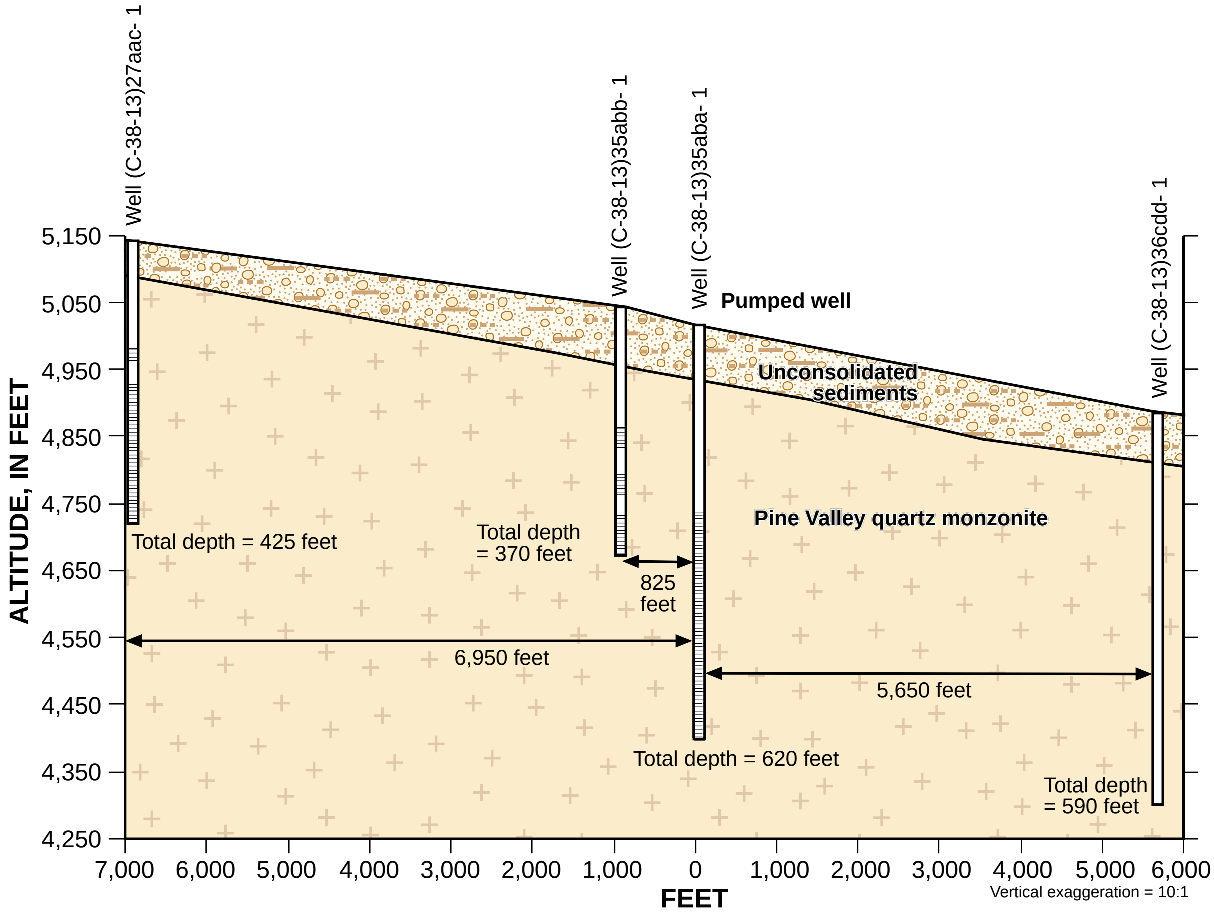


Figure A-19. Schematic cross section of selected wells and lithology of the New Harmony aquifer-test site, Washington County, Utah, October and November 1996.

transmissivity value of 9 ft^2/min (from specific capacity data from the pumped well), but other values of transmissivity and storage did not improve the match.

The drawdown data at the recorder well, however, plots as a straight line with the square root of time as the horizontal axis (fig. A-21). This indicates that the recorder well and the pumped well may be connected by a highly transmissive fracture, which indicates linear rather than radial flow conditions. Jenkins and Prentice (1982) describe: "...an extreme condition where a homogenous aquifer is bisected by a single fracture having a permeability sufficiently large that the ratio of the fracture permeability to the aquifer permeability approaches infinity...Under this extreme condition, flow in the aquifer is linear toward the fracture rather than radial toward the well. (figure A-22) shows a conceptual model of a linear flow system. A homogenous aquifer is bisected by a highly permeable fracture which has been penetrated by a well. When the well is pumped, the water level in the fracture declines, inducing flow into the fracture from the aquifer. The open fracture is a planar production surface that is an extension of the well itself. The well and its hydraulically connected production surface are here called an extended well... The flow lines in the aquifer are parallel; thus, flow in the aquifer is linear and laminar toward the extended well...Drawdown is a function of the perpendicular distance from the extended well, not a function of the radius from the pumped well. Thus, radial flow equations cannot adequately describe aquifer test data from a linear system."

Jenkins and Prentice (1982) also discuss the special case where the observation wells penetrate the production surface of the extended well:

"The drawdown in an observation well which penetrates the production surface of the extended well will be the same as the drawdown in the pumped well, if the pumped well data are corrected for entrance

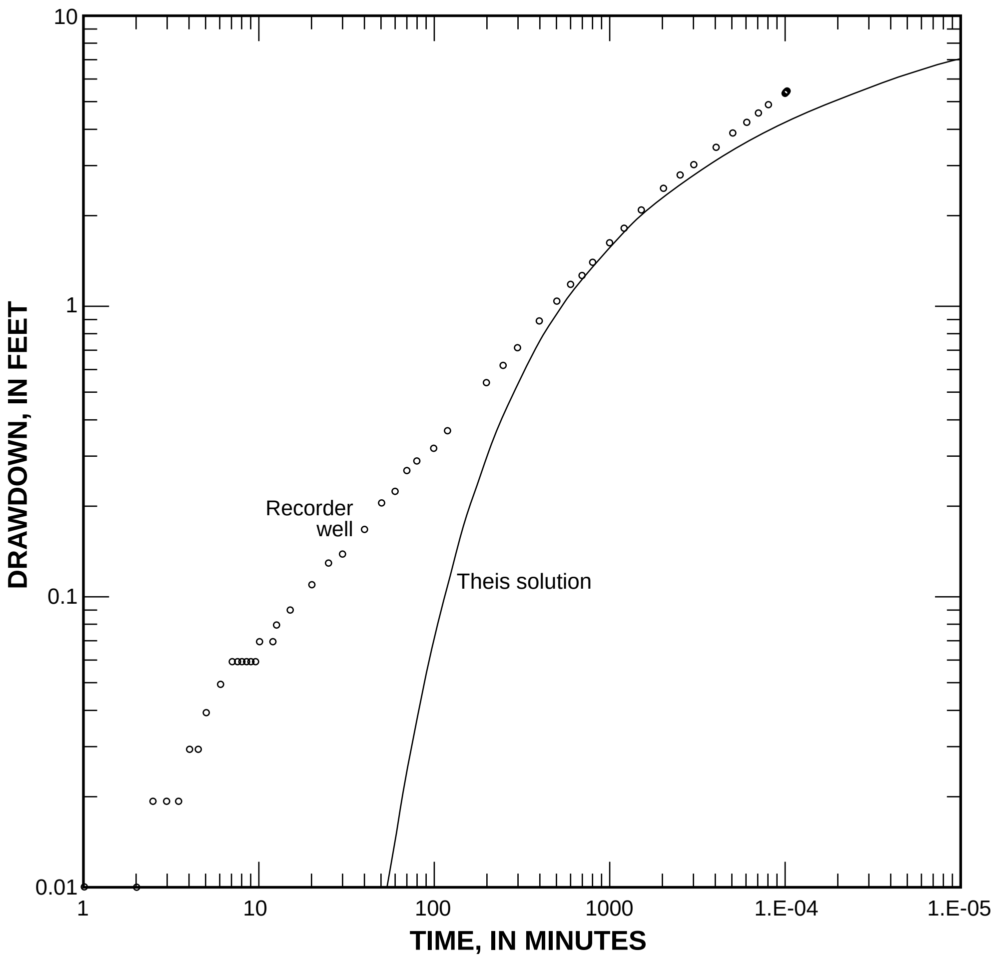


Figure A-20. Drawdown data from the recorder well during the New Harmony aquifer test, Washington County, Utah, October and November 1996 (Unconfined Theis solution).

losses and turbulent flow in the fracture near the well bore. Because the observation well lies along the axis of the trough of depression where water-level declines are greatest...a unique value for T (transmissivity) cannot be determined unless L (fracture length) is known and S (storage) can be reasonably estimated."

Jenkins and Prentice (1982) suggest that flow near the well may be linear if a straight line can be fitted to a plot of drawdown versus the square root of time. Based on this finding, Paul Hsieh (U.S. Geological Survey, written commun., 1997) suggested analyzing the drawdown data of the recorder well using equation 19 of the Jenkins and Prentice paper:

$$s = \left(\frac{Q}{L\sqrt{\pi TS}}\right)\sqrt{t} \qquad \text{(A18)}$$

where s = drawdown,
Q = pumping rate,
L = fracture length,
T = transmissivity,
S = storage, and
t = time.

Equation A-18 is in the form of a linear equation, $y=mx+b$ where the Y axis is drawdown and the X axis if the square root of time. Therefore, the slope of the straight line fitted to the data of figure A-21 is 0.056 and is equal to the expression $\left(\frac{Q}{L\sqrt{\pi TS}}\right)$. Moving the pumping rate *(Q)* and $\pi^{-1/2}$ to the left side of the expression yields the relation: $L\sqrt{TS} = 23.6\,ft^2/(\sqrt{t})$. Because no other information is available to uniquely define storage or transmissivity for the New Harmony aquifer test, this is the quantity reported. Jenkins and Prentice (1982) stated that: "Where L is unknown and the fracture appears to be infinite during an aquifer test, a unique value for T

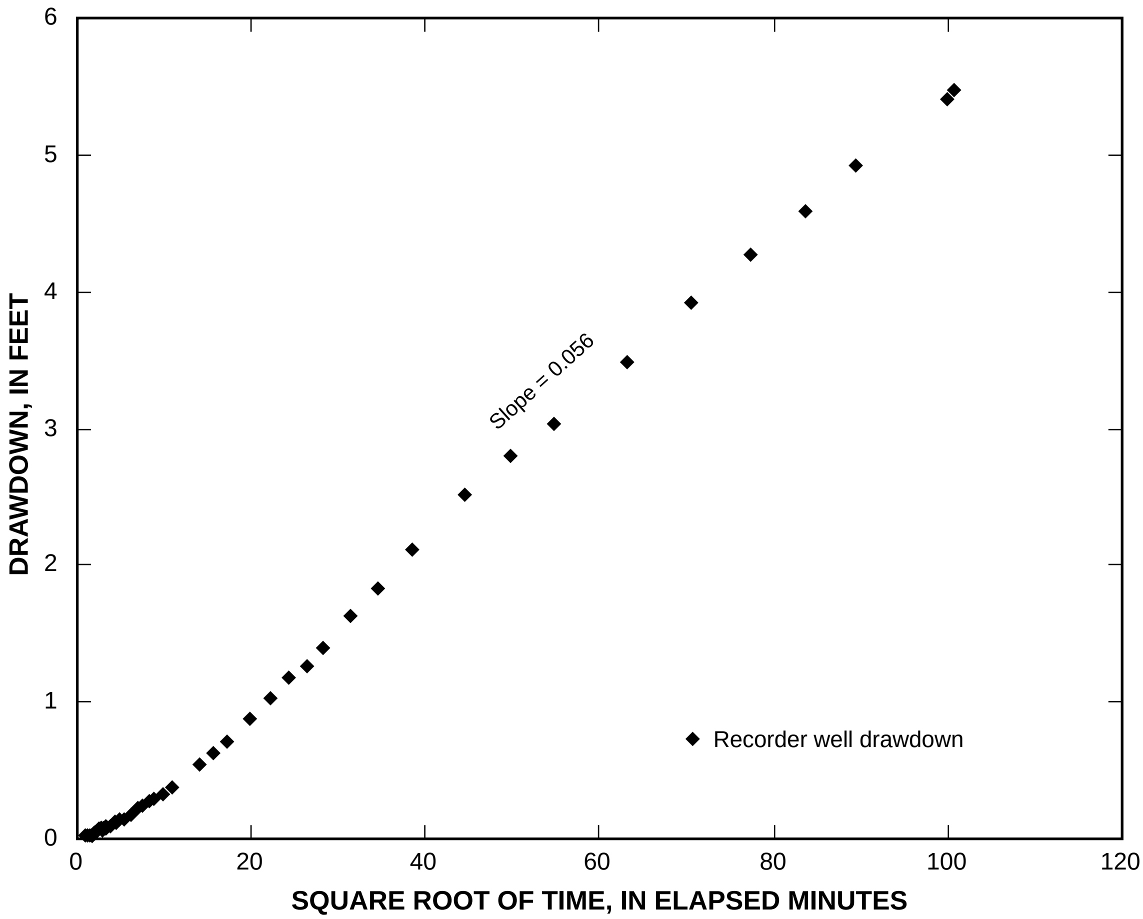


Figure A-21. Drawdown data from the recorder well during the New Harmony aquifer test, Washington County, Utah, October and November 1996 (Jenkins and Prentice solution).

cannot be determined..... Because the length of fractures varies from a few feet to thousands of feet, values of *T* determined using estimated *L* and *S* values should be used with caution.” However, if we assume that the fracture length is at least the 825 ft distance between the pumped well and the recorder well, then the product of *T* and *S* would be

less than or equal to 8.1×10^{-4} ft^2/s.

It should be noted that the aquifer test was only of short duration. Longer-term pumping will result in more drawdown at the pumped well. As stated by Gringarten (1982), the long-term drawdown at the production well can be estimated using the Theis solution with a radial distance half the fracture length. Paul Hsieh (U.S. Geological Survey, written commun., 1997) suggested that “this method of estimating long-term drawdown strongly depends on the estimated fracture length... (Similar to the radial flow case), the drawdown will not stabilize, but is ever increasing, although at a slower and slower rate.”

SUMMARY

Of the 10 observation wells measured during the New Harmony aquifer test, only the recorder well (radial distance of 825 ft) showed substantial drawdown due to pumping. A plot of drawdown data from this well versus the square root of time shows that flow near the well may be linear rather than radial. In a situation where the only affected observation well may intersect the same fracture (or extended well) as the pumped well, a unique value for transmissivity cannot be determined because both the fracture length and storage are unknown. Therefore, the product of fracture length and the square root of transmissivity times storage, $L\sqrt{TS}$,will be reported as about 24 $ft^2/s^{1/2}$.

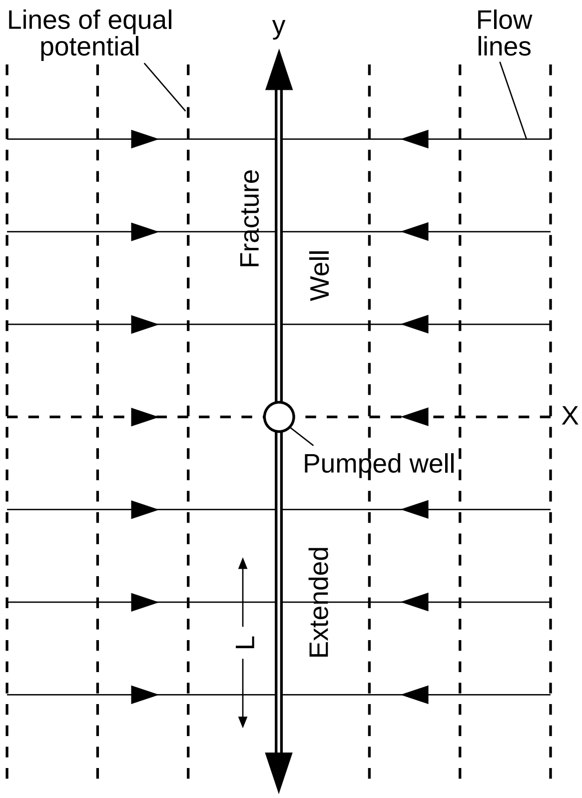


Figure A-22. Conceptual model of a homogeneous aquifer bisected by a single fracture (Jenkins and Prentice, 1982).

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